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The Kalar Anorthosite–Charnokite Complex of the Aldan–Stanovoi Shield: Age and Tectonic Implications

E. B. Sal'nikova*, A. M. Larin*, A. B. Kotov*, V. A. Glebovitskiĭ*, M. K. Sukhanov, S. Z. Yakovleva*, V. P. Kovach*, N. G. Berezhnaya*, and M. D. Tolkachev***

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Abstract—The new data considered characterize age of magmatic rocks from the Kalar Complex of the anorthosite–mangerite–charnokite–granite association (AMCG) in the junction zone of the Aldan shield and Dzhugdzhur–Stanovoi fold system. The U–Pb zircon dating showed that charnokite and gabbro of the complex are the Late Archean in age (2611 ± 38 and 2623 ± 23 Ma respectively) and experienced metamorphism of the granulite facies 1849 ± 15 Ma ago. An earlier granulite metamorphism (2627 ± 16 Ma) that predated the Kalar Complex emplacement was connected with collision between amalgamated terranes, which represent now the granulite basement of the Dzhugdzhur–Stanovoi fold system, and the Olekma–Aldan continental microplate. Accordingly, we suggest that the Kalar AMCG formed during the post-collision time. Based on dating results, we distinguished two AMCG associations of the Aldan–Stanovoi shield, which differ from each other in age and tectonic position. One association corresponds to the Late Archean post-collision plutons of the Kalar Complex, and the other one is represented by the Early Proterozoic (1.74–1.70 Ga) anorthosite massifs of the Dzhugdzhur Complex and by rocks of the Ulkan Complex, which are constituents of the Bilyakchan–Ulkan anorogenic volcano-plutonic belt.

Key words: Early Precambrian, anorthosites, charnockites, geochronology, Aldan shield.

INTRODUCTION

Large anorthosite plutons of the anorthosite–mangerite–charnokite–granite (AMCG) association in terminology of Emslie and Hunt (1990) represent one of geological formations characteristic of ancient cratons. Plutons of this kind originated during the Proterozoic between 1.8 and 1.0 Ga. In the Aldan–Stanovoi shield, there are discriminated two separate distribution areas of the AMCG rocks: the Kalar Complex on the west and the Dzhugdzhur Complex on the east (Fig. 1B). Some researchers (Moshkin and Dagelaĭskaya, 1964; Lennikov, 1979) considered both areas as representing a single East Asian anorthosite belt of the Archean age, which extends in sublatitudinal direction from the Sea of Okhotsk to the eastern Transbaikal region.

The U–Pb zircon date of 1736 ± 6 Ma obtained for anorthosite from the Dzhugdzhur Complex showed that this rock is Early Proterozoic but not Archean in age (Neĭmark *et al.*, 1992; Larin *et al.*, 1997). The earliest attempt to date zircons from the Kalar Complex was that of Levchenkov *et al.* (1987) who selected zircon grains from charnokite but not from anorthosite because of the zircon low abundance in the last rock. The date they obtained (~2.7 Ga) cannot be regarded as correct however, because the analyzed zircons represented a mixture of morphologically different crystals and were selected from rock samples collected from broadly spaced charnokite and enderbite massifs, some of which do not belong to the complex under consideration. There were also attempts to date the Kalar Complex using the Sm–Nd method (Glukhovskiĭ *et al.*, 1993; Sukhanov and Zhuravlev, 2002). The obtained mineral isochrons yielded two concordant (within the analytical uncertainty) age values of 1926 ± 64 and 1913 ± 123 Ma for anorthosite and one more value of 1698 ± 42 Ma for charnokite. Using the Rb–Sr method, Vinogradov *et al.* (1983) who analyzed rock-forming minerals of anorthosite obtained a comparable, though very approximate value of ~1.7 Ga. In view of a considerable discordance between isotopic dates cited above, we decided to date once again different rocks of the Kalar Complex. The new results are discussed below.

GEOLOGICAL BACKGROUND

The Kalar Complex includes the Kuronaakh and Imangakit anorthosite massifs, which are closely spaced, elongated in sublatitudinal direction, and belong to the Kurul'ta block situated in the junction zone of the Aldan shield and Dzhugdzhur–Stanovoi fold system (Fig. 1A). Both massifs represent slab-like

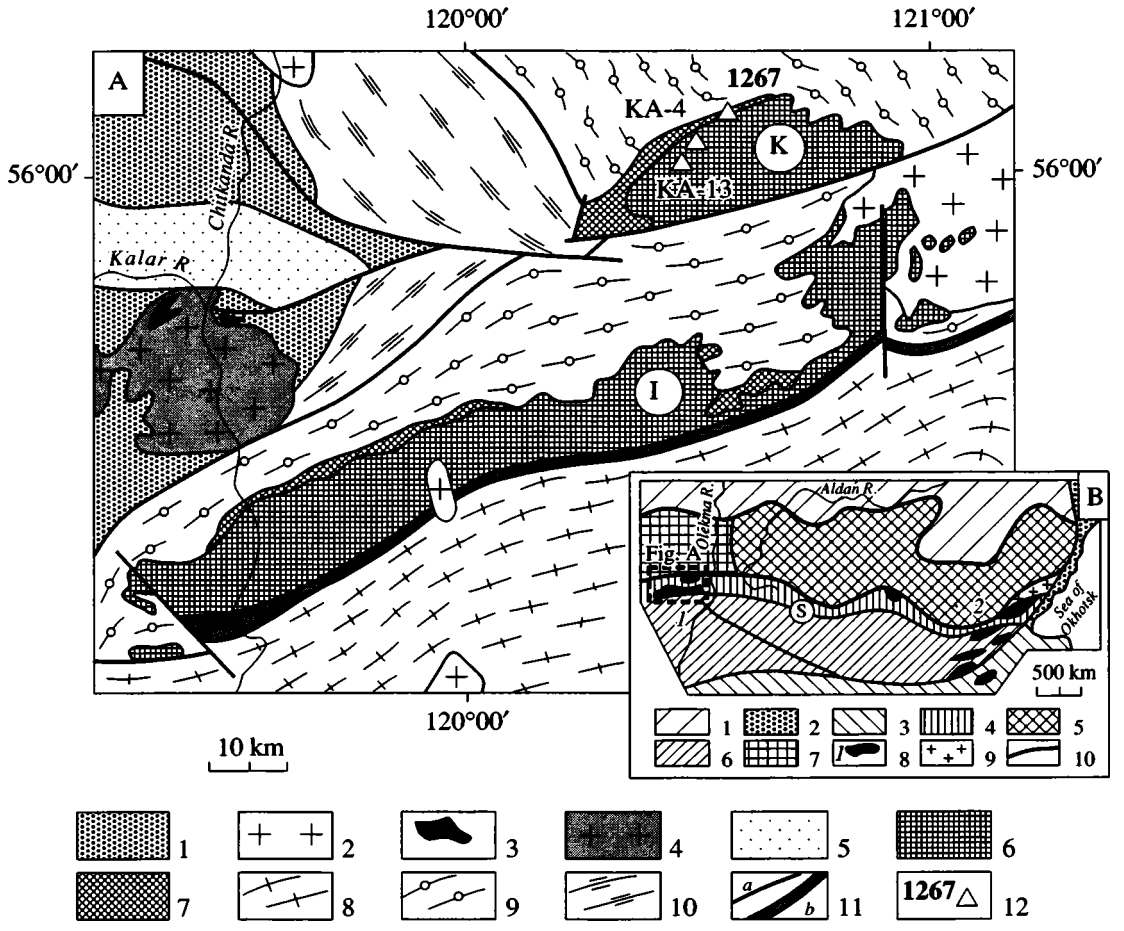

Fig. 1. Schematic geologic map (A) of the Kalar Complex and inset map (B) of regional tectonic units. Legend for A: (1) Phanerozoic sedimentary rocks of superimposed depressions; (2) Paleozoic granite and syenite; (3) Riphean dolerite sills, Doros Complex; (4) granite, Kodar Complex; (5) metasedimentary rocks, Udokan Complex; (6) anorthosite and (7) charnokite of the Kalar Complex; (8) Dzhugdzhur–Stanovoi fold zone; (9) tonalite–trondhjemite gneisses, Olekma Complex; (10) supracrustal enderbites and charnockites of the Kurul'ta block; (11) faults (a) and diaphthoresis zones (b); (12) sampling site. Encircled letters denote the Kuronaakh (K) and Imangakit (I) massifs of the Kalar Complex. Legend for B: (1) sedimentary cover of Siberian platform; (2) Okhotsk–Chukotka volcanic belt; (3) Mongol–Okhotsk foldbelt; (4) high-P granulites of junction zone between Aldan shield and Dzhugdzhur–Stanovoi fold system; (5) Aldan and Batomga blocks of the Aldan shield; (6) Dzhugdzhur–Stanovoi fold system; (7) Chara–Olekma block of the Aldan shield; (8) anorthosite massifs; (9) Ulkan Complex of anorogenic granites and bimodal volcanics; (10) major faults. Designated in the map are the Kalar (*1*) and Dzhugdzhur (*2*) AMCG association and Stanovoi fault (S).

bodies 3 to 4 km thick and 50 to 120 km long, gently dipping to the north. Their contacts with surrounding rocks are tectonic as a rule, although in places there are established the intrusive relationships and high-temperature exomorphism of the Kurul'ta rocks transformed into garnet-pyroxene hornfels (Priyatkina and Lavrovich, 1985). The massifs under consideration are composed predominantly of anorthosite grading near the contacts into melanocratic basic rocks (gabbro-anorthosite, gabbro-norite, and gabbro), which are coarsely layered. These marginal rocks enclose ore bodies composed of apatite, ilmenite, and Ti-magnetite. In the north, where both massifs are not faulted against the country rocks, they are bounded by charnokite, the outer contact of which is distinctly intrusive. In general, relationships between anorthosite, gabbroic rocks, and charnokite are interpreted controversially, although the common opinion is that these rocks are interrelated via zones of gradual transition (Priyatkina and Lavrovich, 1985; Sukhanov and Vaganov, 1991).

Geological characters of anorthosite, gabbro, and charnockite, as well as their mineral and geochemical peculiarities (Sukhanov and Vaganov, 1991; Glukhovskiĭ *et al.*, 1993; Larin and Rundkvist, 1999; Mints *et al.*, 2000; Sukhanov and Zhuravlev, 2002), imply that all these rocks belong to the AMCG association. Dominant in the charnokite series are orthoclase granitoids with augite, hypersthene, hastingsite, and fayalite. According to geochemical data, these are intraplate and A-type granitoids. It is remarkable as well that, in contrast to the Dzhugdzhur Complex, rocks of the Kalar Complex were affected by granulite metamorphism and afterward experienced diaphthoresis.

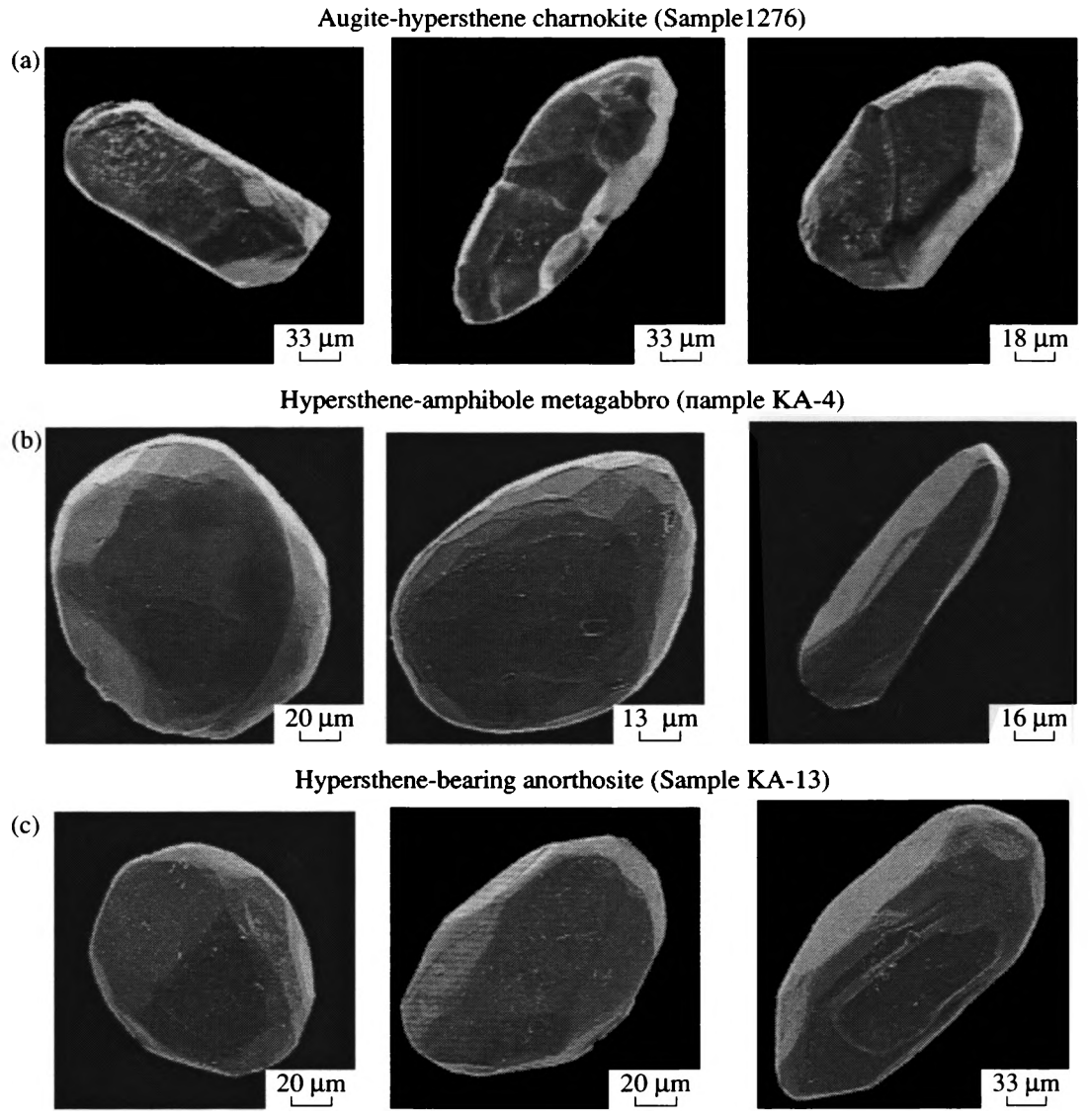

Fig. 2. Microphotographs of zircons from augite charnokite (a), hypersthene-amphibole metagabbro (b), and hypersthene-bearing anorthosite (c) of the Kalar Complex (scanning electron microscope ABT-55, accelerating voltage 20 kV).

ANALYTICAL METHODS

We used the standard method of zircon separation in heavy liquids. Subjected to the U–Pb isotopic analysis were both the bulk zircon separates (10 to 50 grains) and individual zircon crystals preliminary cleaned of impurities with the help of aeroabrasion (Krogh, 1982). The internal structure of individual crystals was studied under optical microscope and by cathodoluminescence (CLC) method (Poller *et al.*, 1997). When the latter procedure was applied, we used approximately a half of the examined zircon crystal directly for the U–Pb isotopic analysis. Any selected crystal was treated then in alcohol, acetone, and 1M HNO_3 in order to remove surficial impurities, with each step followed by rinsing in extrapure water. The modified chemical method of Krogh (1973) was used for decomposition of crystals and extraction of U and Pb for subsequent analysis. The blank run during the analytical period was not greater than 30 pg Pb. Isotope composition of U and Pb was determined on the multi-collector Finnigan MAT 261 mass spectrometer using a regime of simultaneous measurements of all isotope currents or electronic multiplier unit (discrimination coefficient of 0.32 ± 0.11 amu for Pb). The measurement results are processed using geochronological toolkits "PbDAT" (Ludwig, 1991) and "ISOPLOT" (Ludwig, 1999), and applying the standard U decay constants (Steiger and Jager, 1976). Corrections for common lead correspond to those accepted in model by Stacey and Kramers (1975). All the uncertainties are calculated at the 2σ level.

Cathodoluminescence of zircon crystals was examined under scanning electron microscope "CamScan

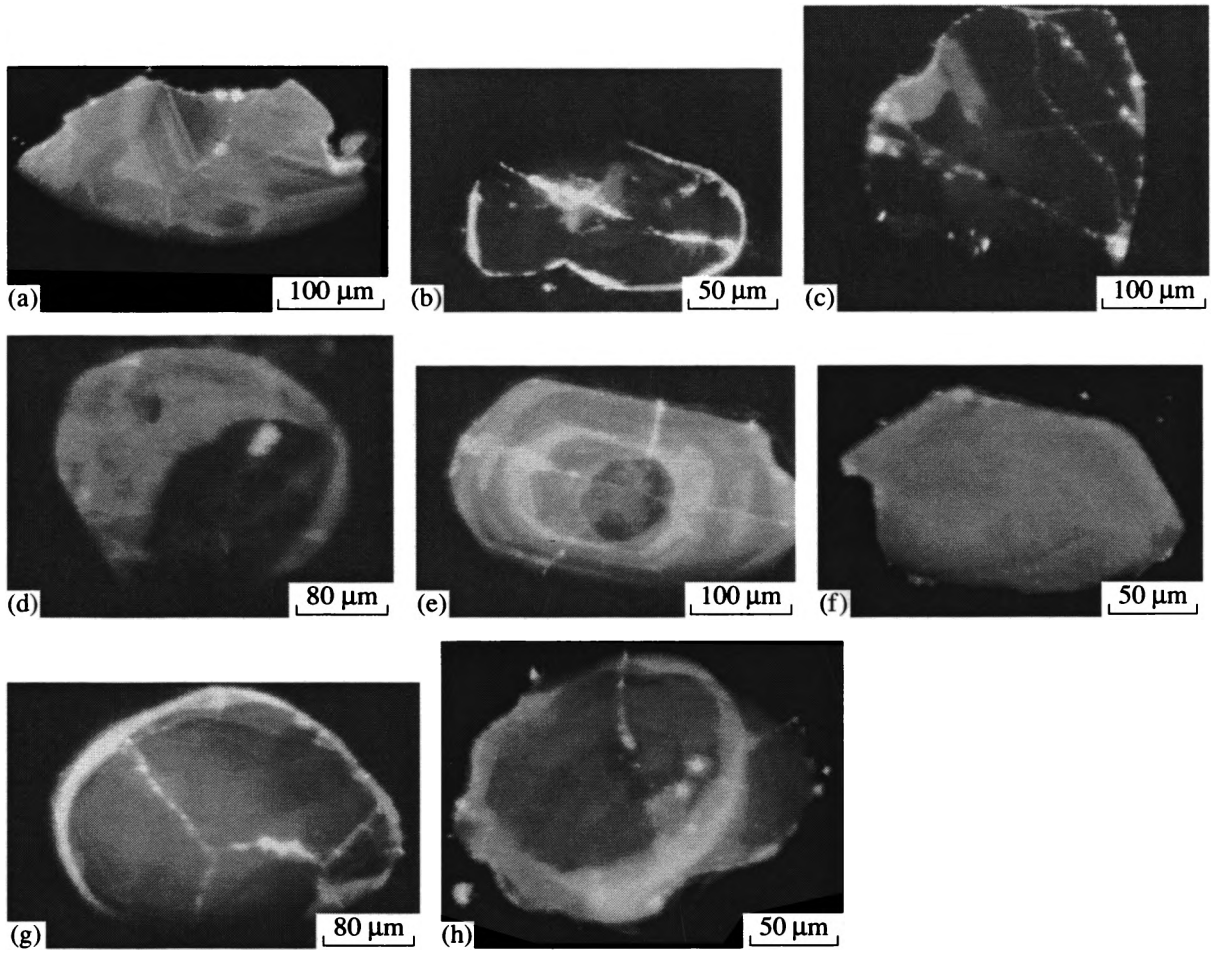

Fig. 3. Microphotographs of zircons from hypersthene-amphibole metagabbro, Sample KA-4 (a–d), and hypersthene-bearing anorthosite, Sample KA-13 (e–h), of the Kalar Complex (scanning electron microscope "CamScan" equipped with cathodoluminescence detector "Centaurus," accelerating voltage 15kV); crystals figured in (a) and (b) have been selected for isotopic analysis (lines 8 and 9 in the table).

MV2300" equipped with detector "Centaurus." The optimal working length was 12 to 15 mm, current beam intensity not less than 1 nA and accelerating voltage of electron gun 15 KV. Intensifier "BIALKALI" with bandwidth of 300–650 nm in visible spectrum enabled visualization and digital recording of resultant CL images of zircon crystals.

GEOCHRONOLOGICAL RESULTS

Three samples of different rock types suitable for geochronological analysis have been collected from the Kuronaakh massif. These are augite–hypersthene charnokite (quartz monzonite, Sample 1267), hypersthene–hornblende gabbro from the massif marginal zone (Sample KA-4), and hypersthene-bearing anorthosite (Sample KA-13). We dated anorthosite and associated charnokite, since it was necessary to define whether they are cogenetic, i.e., belonging to one and the same AMCG association, or not.

Augite charnokite. Accessory zircons from this rock are represented by transparent and translucent crystals of subidiomorphic prismatic and long-prismatic habit (Fig. 2a), pinkish lilac in color. Characteristic of the crystals are zoned cores surrounded by thin unzoned outer rim. Crystals range in size from 50 to 600 μm having elongation coefficients 2.2 to 5.5.

Three groups of zircon crystals have been selected from size fractions <150 and >150 μm for the U–Pb isotopic analysis. Before the analysis, crystals of coarser fraction were subjected to aeroabrasion cleaning (see lines 2 and 3 in the table). Data points corresponding to measurement results plot in Fig. 4a along a discordia, the upper intercept of which with concordia corresponds to 2611 ± 38 Ma and the lower one to 1668 ± 40 Ma (MSWD = 0.01).

Morphology and inner structure of zircon crystals from augite–hypersthene charnokite suggest their magmatic origin and subsequent transformations in the course of high-temperature metamorphism. We think

U–Pb isotopic data on zircons from the Kalar Complex

No.	Fraction size (μm) and characterization	Weight (mg)	Concentration (ppb)		Isotopic ratios						Rho	Age, Ma		
			Pb	U	Th/U[a]	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$[b]	$^{208}Pb/^{206}Pb$[b]	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$		$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$
Augite charnokite (Sample 1276)														
1	<150	0.53	41.7	87	0.48	2199	0.1580 ± 1	0.1329 ± 1	9.3410 ± 187	0.4287 ± 9	0.83	2372 ± 5	2300 ± 5	2435 ± 1.2
2	>150 A 50%	0.45	22.6	43	0.51	1106	0.1651 ± 1	0.1407 ± 1	10.352 ±21	0.4548 ± 9	0.89	2467 ± 5	2116 ± 5	2509 ± 1.2
3	>150 A 70%	0.24	39.0	75	0.48	798	0.1633 ± 1	0.1335 ± 1	10.072 ± 20	0.4474 ± 9	0.85	2441 ± 5	2384 ± 5	2490 ± 1.2
Hypersthene–amphibole metagabbro (Sample KA-4)														
4	–100 + 80	0.48	88.3	173	0.72	3935	0.1594 ± 1	0.1984 ± 1	9.5829 ± 192	0.4360 ± 9	0.93	2396 ± 5	2333 ± 5	2449 ± 0.6
5	>100	0.88	48.8	90	0.79	5136	0.1657 ± 1	0.2175 ± 1	10.469 ± 20	0.4581 ± 9	0.93	2477 ± 5	2431 ± 5	2515 ± 0.6
6	>100, R	–	U/Pb* = 1.9		0.87	13092	0.1524 ± 1	0.2411 ± 1	8.9462 ± 179	0.4258 ± 9	0.97	2333 ± 5	2287 ± 5	2373 ± 0.5
7	>75, A 20%	0.47	42.1	84	0.79	3708	0.1546 ± 1	0.2192 ± 1	9.0155 ± 180	0.4229 ± 8	0.94	2340 ± 5	2274 ± 5	2398 ± 0.8
8	IC (a, Fig. 3)	–	U/Pb* = 2.0		0.72	904	0.1538 ± 2	0.1990 ± 1	8.9748 ± 404	0.4232 ± 18	0.94	2335 ± 11	2275 ± 10	2389 ± 2.5
9	IC (b, Fig. 3)	–	U/Pb* = 2.0		0.18	803	0.1673 ± 2	0.0493 ± 1	10.696 ±27	0.4645 ± 11	0.93	2497 ± 6	2459 ± 6	2528 ± 1.5
Hypersthene-bearing anorthosite (Sample KA-13)														
10	<80	0.55	108	299	0.53	5387	0.1143 ± 1	0.1519 ± 1	5.1766 ± 104	0.3286 ± 7	0.95	1849 ± 4	1832 ± 4	1868 ± 0.8
11	>100	1.05	78.2	168	0.91	18892	0.1397 ± 1	0.2622 ± 1	7.4130 ± 148	0.3850 ± 8	0.97	2163 ± 4	2100 ± 4	2223 ± 0.6
12	>150	0.63	86.7	202	0.83	3713	0.1279 ± 1	0.2384 ± 1	6.3728 ± 127	0.3615 ± 7	0.97	2029 ± 4	1989 ±4	2069 ± 0.8
13	<80, R 3 h	0.82	68.3	177	0.72	8866	0.1161 ± 1	0.2056 ± 1	5.3974 ±108	0.3372 ± 7	0.96	1884 ± 4	1873 ±4	1897 ± 0.6
14	>100, R 3.5 h	–	U/Pb* = 2.5		0.72	24802	0.1216 ± 1	0.2061 ± 1	5.8196 ±116	0.3472 ± 7	0.92	1949 ± 4	1921 ±4	1979 ± 0.9

Note: (a) calculated based on $^{208}Pb/^{206}Pb$ ratio and $^{207}Pb/^{206}Pb$ age value; (b) ratio corrected for blank and common leads; (A 50%) percentage of abraded material; (R 3h) residue after acid leaching during 3 hours; (*) zircon fraction has not been weighted; (IC) individual crystal selected after cathodoluminescence examination. Uncertainties correspond to 2 σ.

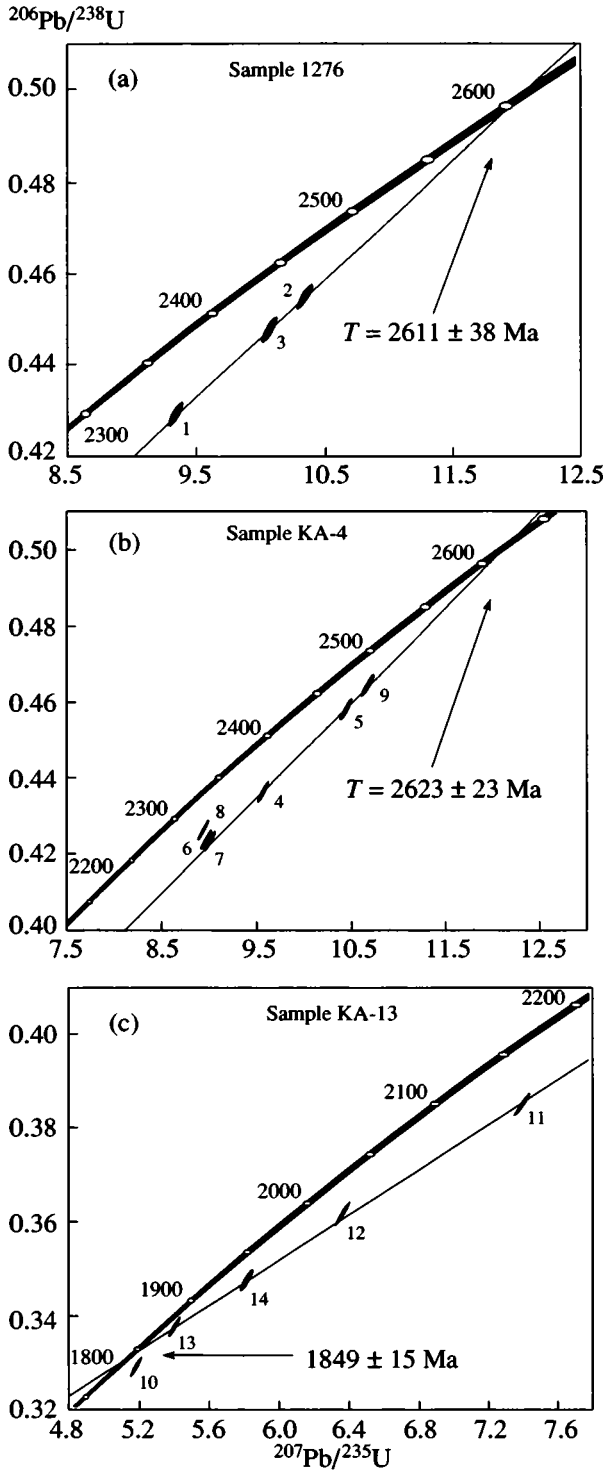

Fig. 4. Concordia diagrams for zircons from augite charnokite (a), hypersthene-amphibole metagabbro (b), and hypersthene-bearing anorthosite (c) of the Kalar Complex.

that all secondary phases of crystals were eliminated during the abrasive cleaning, and the age of the upper discordia intercept characterizes therefore the crystallization time of parental charnokite magma.

Hypersthene–amphibole metagabbro. This rock yielded zircons of the following types: (1)transparent and translucent subidiomorphic and rarer idiomorphic multiplanar crystals of isomorphic shape; (2) irregularly faceted crystals approaching the isometric shape and partially retaining facets of bipyramid; and (3) crystals of prismatic habit (Fig. 2b). Coloration of the crystals changes from pinkish to dark cherry-red, and their size ranges from 30 to 200 μm, elongation coefficient 1.0 to 2.0. The CLC examination showed that crystals have "magmatic" cores with trace zonation and outer unzoned rims with a bright luminescence (Fig. 3). In some cores, there are recognizable signs of recrystallization under high-temperature conditions.

At the beginning, we investigated three groups of dark red zircon crystals from size fractions of 100–80, >100, and <75 μm, which were more transparent, than the other grains (see lines 4, 5, and 7 in the table). All these zircons, including those subjected to abrasive cleaning, turned out to be essentially discordant (Fig. 4b). To decrease the discordance degree, we used the acid dissolution technique (Mattinson, 1994) to clean the larger crystals. The data point characterizing isotopic composition in residues (line 6 in the table) is displaced to the left of primary discordia ($^{207}Pb/^{206}Pb$ age = 2373 ± 0.5 Ma; see Fig. 4b). It is likely that the acid treatment removed the disturbed zircon phases not only from younger outer shells, but from the crystal cores as well, and the analyzed residues represented therefore a mixture of asynchronous, presumably Late Archean and Early Proterozoic crystalline phases. At the next step of investigation, we analyzed individual zircon grains selected directly from the CLC preparations (Poller *et al.*, 1997), cathodoluminescence of which was preliminary examined. The isotopic composition of zircon crystal that showed a weak luminescence (Fig. 3b) corresponds to a data point on the primary discordia and to the minimal discordance of U/Pb ratios (Fig. 4b; line 9 in the table). In contrast, the crystal with a brighter luminescence (Fig. 3a) is most discordant (Fig. 4b; line 8 in the table). Its $^{207}Pb/^{206}Pb$ age of 2389 ± 2.5 Ma is most likely a consequence of zircon recrystallization in the course of superimposed metamorphism. The discordia calculated for three zircon fractions and one individual crystal (Fig. 4b; line 9 in the table) has upper and lower intercepts with concordia, which correspond to age values of 2623 ± 23 and 1776 ± 68 Ma, respectively (MSWD = 1.12). As the discordia is plotted for magmatic zircons practically lacking signs of recrystallization under high temperature, we can regard the age estimated for the upper discordia intercept (2623 ± 23 Ma) as corresponding to crystallization time of parental gabbroic magma.

Hypersthene-bearing anorthosite. Transparent and translucent subidiomorphic zircon crystals from anorthosite are oval, isometric and rarely prismatic in shape and pinkish to cherry-red in coloration (Fig. 2c). The CLC method showed presence of dark cores surrounded by a wide zoned envelope with a bright lumi-

nescence (Fig. 3). Crystals range in size from 30 to 400 μm, elongation coefficient 1.0 to 3.0. Isotopic analysis is performed for five zircon separates from size fractions of <80, >100, and >150 μm (the table), two of which were subjected to preliminary acid cleaning (lines 13 and 14 in the table). As one can see from Fig. 4c, data points characterizing coarse zircon fractions and residue after acid leaching of fine fraction (<80 μm) determine a discordia and plot close to its lower intercept with concordia defining the age value of 1849 ± 15 Ma; the upper intercept corresponds to the age value of 2651 ± 54 Ma (MSWD = 1.3).

Morphology and inner structure of zircons from metagabbro and anorthosite suggest that crystals were transformed (recrystallized and surrounded by metamorphic envelopes) during the granulite metamorphism. The most altered are zircons in anorthosite. That metamorphic event also resulted in formation of a new zircon generation in anorthosite. The lower intercept of discordia plotted for zircons from anorthosite (Fig. 4c) implies that granulite metamorphism in question took place 1849 ± 15 Ma ago.

CONCLUSION

The dating results evidence in favor of cogenetic origin and Late Archean age of anorthosite, gabbro, and charnokite from the Kalar Complex. Thus, it is possible to state that magmatic rocks of the complex belong to one AMCG association which is the most ancient one among its known representatives. The granulite metamorphism that predated emplacement of the Kalar Complex (2623 ± 23 Ma) resulted in formation of charnokite–enderbite plutons 2627 ± 16 Ma ago (Kotov, 2003). This metamorphic event was connected with collision between amalgamated terranes, which represent now the granulite basement of the Dzhugdzhur–Stanovoi fold system, and the Olekma–Aldan continental microplate (Kotov, 2003; Larin *et al.*, in press). The time span separating the collision event from emplacement of massifs of the Kalar AMCG association was about 30 m.y. long, as we can judge from the estimated age values with due account for their uncertainty. Accordingly, it is admissible to think that the Kalar Complex originated in the post-collision setting. It is also remarkable that between 2.68 and 2.60 Ga there were several pulses of post-collision granitoid magmatism in the southwestern margin of the Aldan shield (Rizvanova *et al.*, 1994; Kotov *et al.*, 1995; Larin *et al.*, 2002a; Kotov, 2003). Presumable geological analogues of the Kalar Complex are the AMCG plutons (1.16–1.13 and 1.09–1.05 Ma old) of the Grenville orogen, which are also of the post-collision type (Corrigan and Hanmer, 1997). Emplacement of the latter is explained in terms of a model of lithosphere delamination, subsequent uplift, and extension collapse of the collision orogen. Associated with these events were intrusions of high-temperature mafic magmas generated in asthenosphere.

The high-grade metamorphism (~1.9 Ma ago) that is recorded in rocks of the Kalar Complex was most likely related with docking of the Dzhugdzhur–Stanovoi continental microplate to the Siberian craton (Mints *et al.*, 2000; Larin *et al.*, 2002b; Larin *et al.*, in press) after closure of former oceanic structure in the Early Proterozoic. The trace of that structure is probably the Stanovoi structural suture separating the Dzhugdzhur–Stanovoi fold system from the Aldan shield (Gusev and Khain, 1995; Mints *et al.*, 2000). The high-grade metamorphism of that time is recorded in granulite blocks in the Dzhugdzhur–Stanovoi folded zone and Stanovoi structural suture. The syncollision magmatism resulted in formation of granitoid plutons (diapirs) of the Kuanda Complex, which are 1895 ± 30 Ma old (Rublev *et al.*, 1987), confined exclusively to the southern foreland of the Aldan shield. The postcollision magmatism in the same marginal zone produced large granite massifs of the Kodar Complex within a narrow time span 1876–1873 Ma ago (Larin *et al.*, 2000).

The Sm–Nd age value of 1698 ± 42 Ma that has been obtained earlier for charnokite of the Kalar Complex (Glukhovskiĭ *et al.*, 1993; Sukhanov and Zhuravlev, 2002) is concordant, within the uncertainty limits, with age values at the lower discordia intercepts, which are calculated for charnokite (1668 ± 40 Ma) and metagabbro (1776 ± 68 Ma) in this work. All these values are close to the time (1.74–1.70 Ga) of the last Early Precambrian anorogenic event in the Aldan–Stanovoi shield (Larin *et al.*, 1997), when considerable vertical movements displaced blocks of the platform basement. Exhumation of the Kurul'ta tectonic block probably happened in this period in connection with tectonic activation of the Stanovoi structural suture. An argument in favor of this inference is the K–Ar age of 1655 ± 35 Ma obtained for phlogopite from pegmatite of that block (data of M.M. Arakelyants, IGEM RAS).

Thus, two AMCG associations different in age and tectonic significance are distinguishable in the Aldan–Stanovoi shield. The post-collision plutons of the Late Archean Kalar Complex are localized in the Kurul'ta block of junction zone between the Aldan shield and Dzhugdzhur–Stanovoi fold system, whereas the Dzhugdzhur Anorthosite and Ulkan Complex of the Early Proterozoic (1.74–1.70 Ga) AMCG association belong to the anorogenic Bilyakchan–Ulkan volcanoplutonic belt. This belt extending for a distance over 750 km from the Mongol–Okhotsk suture to the Okhotsk massif links the Early Precambrian tectonic structures in the southeastern Siberian platform (Larin *et al.*, 1997, 2002a).

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Evolution of Phanerozoic Miospores and Main Striation Types

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Abstract—Morphogenetic peculiarities of exine striation types characteristic of the Phanerozoic miospores are discussed. During the period over 400 m.y. long, terrestrial plants produced 21 types of miospores with variably striated exine. Most of them originated in the Devonian and Permian epochs, while afterward the striation development decelerated so that younger forms are characterized in general by previous striation patterns. Among Triassic striate miospores there are two new types only, whereas Jurassic and Cretaceous epochs produced one new type each. Representatives of the Paleogene, Neogene–Quaternary, and Recent miospores inherit the striation types characteristic of older taxa. The evolution of striate miospore was of the iterative type, as it follows from changes in disposition of striate sculptural elements. The evolution of land vegetation was accompanied by recurrent development of some striation types in the course of appearance of new plant and miospore taxa. Main morphotypes of striate miospores and their time succession are described. Origination and extinction of different striate morphotypes, which are confined to certain stratigraphic units, are recorded in a strict chronological succession. Type species of striate miospores typical of main Phanerozoic stratigraphic units and their stratigraphic ranges are characterized.

Key words: Miospores, striation types, exine ornamentation, palynoflora, recurrence, Phanerozoic.

INTRODUCTION

This study continues a series of works dedicated to evolution of miospore striation during the Phanerozoic (Podgainaya, 1993, 1995, 1996a, 1996b, 1999, 2000).

The study of striate miospores and simultaneous definition of the striation notion commenced synchronously with development of palynology as an autonomous science. Jackson (1928) was first to define the striation notion and referred miospores with fine longitudinal elements (furrows and ridges) to striated varieties. He related the cell wall striation to the protoplasm activity. Potonie (1934) considered miospores with costate exine ornamented with furrows and similarly wide or narrower dividing ridges as striated. According to Faegri and Iversen (1950), striate miospores include pollen grains with threads or rugulate ornamentation in exine. In opinion of Erdtman (1952), streaked sculptural thickenings of exine originate by horizontal fusion of tubercles (or their tops) arranged in chains. Usually, the ornamentation in question is represented by narrow parallel grooves and ribs twice longer than their width. Thompson and Pflug (1953) referred to striate miospores the pollen grains with elongated structural elements arranged in a parallel, radial, or some other manner. Harris (1955) argued that pollen grains with regular distinct ribs and grooves covering their surfaces should be referred to costate miospores. Kupriyanov and Aleshina (1972, 1978) considered grains with meridional parallel streaks, striae, or costae in their exine as representing the true striate miospore.

Kremp (1967) proposed the most complete and comprehensive definition of the striation notion, according to which striate miospore are the pollen grains with taeniate, grooved, cristate, streaked, canaliculate, cicatricose, costate, rugulate–striate, and striate–rugulate ornamentation in the exine. Relief-forming sculptural elements of the exine in such miospores are represented by rollers, costae, ridges, furrows, and striae arranged uniformly at the exine surface. In addition, the group includes plicate and polyplicate miospores, in which meridional furrows and striae resulted from the entire exine contraction in a certain direction or are formed by internal, not surficial sculptural elements of the exine.

Thus, most of the researchers consider the relief-forming elements of the exine as striate ornamentation, almost ignoring peculiarities in their arrangement on the exine surface.

Similarly Meyen (1987), who summarized data on morphologic evolution of Phanerozoic miospores and described geochronological succession of their morphotypes, apertures, and exine sculpturing, paid attention to morphogenetic peculiarities of several striate genera only: *Vittatina, Weylandites, Gnetaceaepollenites, Ephedripites*, and *Classopollis*. I reviewed a great number of palynological publications (over 500 titles) dedicated to morphological structure and stratigraphic distribution of individual striate miospore species. The review showed that the following aspects of striation phenomenon are left almost without attention of researchers: (1) historical–genetic succession of stria-

tion types and stratigraphic confinement of changes in striate ornamentation; (2) classification of striate elements arrangements; and (3) general trends in evolution of miospores striation during the Phanerozoic.

In this work, the attention is paid to these problems exactly. My own long-lasted observations on palynofloras from the Upper Paleozoic, Mesozoic, and Cenozoic sediments and data published on the issue represent the factual base of the work. The proposed classification of striation types should be considered as a preliminary version that can be modified and extended when new data are obtained.

RESULTS AND DISCUSSION

The first striate miospores appeared at the Silurian–Devonian boundary time simultaneously with appearance of terrestrial higher plants (Podgainaya, 1993). The striation of miospores represents a protective adaptation mean and fulfills the harmomegathi function that regulates the miospore volume under contrasting temperature and moisture conditions to preserve the reproductive cells in extreme or close to extreme environments.

The number of "striated genera" (Figs. 1, 2) was different in each geological period during the over 400-m.y.-long development of striation phenomenon (Podgainaya, 1995, 1996, 2000). In addition to changes in abundance of "striated taxa," the morphological structure and arrangement of striate sculptural elements also varied substantially (Figs. 1, 2; Appendix to Fig. 2; Plate). These transformations are considered below per each geological period separately.

In the **Devonian**, 18 form genera of striate miospores are registered. In the terminal Silurian–initial Devonian, spherical spores of the genus *Emphanisporites* Mc Greg. with radiating ribs on the proximal face were of a global distribution (Fig. 1, fig. *1*, Plate, fig. *1*). Representatives of the genus have ribs regularly thickening from the spore pole toward equator. Some researchers (Mc Gregor, 1984, Plate 6, fig. 9; Arkhangel'skaya, 1985, Plate 4, fig. 15; Richardson and Mc Gregor, 1986, Plate 2, fig. 12) also reported on some forms and species with ribs radiating away from equator and with pole areas free of ornamentation (*E. protophanus* Richard. et Ioanides, Fig. 1, fig. *2*). There were also species with stellate-intraradial muri in interradial sectors of the spore proximal face (*E. schultzii* (Schultz) Mc Greg., Fig. 1, fig. *3*, Plate, fig. *2*).

Origination of these simple striation types was followed by a tendency of sculptural elements complication and fusion of different exine projections and internal structural elements into striae. For example, structural elements of this kind are represented by proximal submeridional vermiculate–discrete ribs of *Rugosospora flexuosa* (Jusch.) Streel and by radial infrastructure in *Archaeozonotriletes chulus* Cramer var. *inframurinatus* Richard. et Lister. Remarkable are radial structures of similar width between the pole and equator of the spore.

The Early–Middle Devonian boundary period is marked by origination of a new arrangement style of striate sculpturing elements: tubercles fused into vermiculate ribs arranged in the triangular–parallel manner (*Clivosispora verrucata* var. *convoluta* Mc Greg. et Camf. (Fig. 1, fig. *7*)

Several new striation types appeared in the **Middle Devonian.** These are thick roller-shaped structures chaotically interlaced at the distal and proximal faces in spores of the genus *Periplecotriletes* (Naum.) Arch., or formed by chaotically arranged tubercles located on wide perisporium in *Archaeozonotriletes polymorphus* Naum. (Fig. 1, figs. *6*, *9*). In spores *Retusotriletes striatus* (Peppers (Pep. et Domb.), the proximal face is covered by the striated sculpture formed by fused granules of curvatura (Fig. 1, fig. *5*).

In the **Middle–Late Devonian**, most spores retained the radial arrangement of striate sculptural elements, although it became more diverse than before. Thin radial muri are typical of *Emphanizonotriletes solantaicus* (Arch.) Arch. and *Diducites radiatus* (Kedo) Obuch. Spores of *Geminospora bislimbata* (Tschibr.) Mc Greg. et Camf. are characterized by radial vermiculate muri located on the wide perisporium. Sometimes, striation is represented by radial thin filiform or platy structures (*Lanatisporites bislimbatus* (Tschibr.) Arch.), or by thin rods fused into radial bands (*Membrabaculisporites radiatus* (Naum.) Arch.) and by radial costae diverging from the center (*Archaeotriletes solaris* Nekr.). In rounded–triangular spores of *Hymenozonotriletes radiatus* Naum., radial plicae of exine extend from the proximal face to the filmy perisporium (Fig. 1, fig. *4*). The monolete spores *Limbomonoletes lenis* Arch. have longitudinal–parallel striation formed by exine thickenings parallel to the longitudinal axis of the bilateral spore (Fig. 1, fig. *10*).

Naumova (1953) discovered rounded aperture-free palynomorph *Chomotriletes vedugensis* Naum. ornamented with thin concentric–circular costae in the Upper Devonian (Frasnian) sediments of the Russian platform (Fig. 1, fig. *8*). Concentric muri are arranged densely like in a spire, and I term this type of striate ornamentation as spiral. The biological affinity of this palynomorph is problematic and its reference to spores is questioned. Because of its ambiguous biological origin, I share the opinion of Naumova, taking into consideration the palynomorph *Chomotriletes multivittatus* Playford found in the Lower Carboniferous Ducabrook Formation of eastern Australia (Playford, 1978, p. 144, Plate 13, figs. 1–10), which is similar in morphology to the above form.

Thus, nine arrangement types of striate sculptural elements originated in the Devonian: radial-expanding, radial-equatorial, radial, intraradial, spiral, chaotic, longitudinal-parallel, triangular-parallel, and streaked. All

these types of exine sculpturing are characteristic of morphology of later miospores (Fig. 2).

The **Carboniferous** palynofloras include 36 striate form genera. Spores with radial striation still occur in the Early Carboniferous, although the arrangement of their striate elements is more diverse. For example, radial furrows in *Hymenospora caperata* Fekix et Burbridge are located on the filmy perisporium. The radial striation is also typical of *Densosporites striatiferus* High et Playf. and *Radiizonates striatus* (Knox) Stapl. et Jans, (Fig. 1, fig. *5*). *Cirratriradites radialis* (Jusch. et Byvsch.) Byvsch. bears the radiate-rugulate zone. In *Archaeozonotriletes intrastriatus* Playf. (Plate, fig. *3*), striation consists of fine radial dichotomizing striae. In *Radialetes costatus* Playf., the entire surface of miospore is covered by radial muri. At the distal face of *Rugospora radiata* (Kedo) Byvsch., ectexine is covered by radial vermiculate and discrete muri. *Stellispora actinomorpha* (Byvsch.) Byvsch. is characterized by small radial cristae at the distal face, which are stellate in intraradial areas of the proximal face.

In addition to representatives of the genus *Periplecotriletes* (Naum.) Arch., chaotic striation type is characteristic of other genera and species as well. For example, muri chaotically interlaced on the body of *Velamisporites irrugatus* Playf. continue to its zone also (Fig. 1, fig. *2*). In the genus *Cincturasporites* (Hacg et Barrs) Byvsch., the distal surface of miospore bodies is covered with chaotically arranged muri extending to the cingulum (Fig. 1, fig. *3*). The chaotically arranged muri cover the distal surface of miospores *Arcuatisporites* Tet. occurring in the Lower–Middle Carboniferous boundary strata.

The **Middle Carboniferous** was the appearance time of genera *Proprisporites* Neves and *Vestispora* Wils. et. Venk (Fig. 1, figs. *8, 9*, Plate, fig. *4, 5*), which include miospores with a glomerate arrangement of sculpturing elements and with two layers of superimposed thin muri at least observable on the spores surface. The wide striation elements of spores *Concentricisporites concentricus* Byvsch. form several (3–5) concentric circles on the spore body (Fig. fig. *7*). There are also miospores with spiral arrangement of numerous thin muri (*Chomotriletes multivittatus* Playf., Fig. 1, fig. *6*) and with triangular-parallel arrangement of vermiculate sculptural elements repeating the equatorial contour of spores (*Callisporites* Butt. et Will, Fig. 1, fig. *4*). The longitudinal-parallel striation is characteristic of monolete miospores *Laevigatosporites costatus* Alp. and *Columinisporites heyleri* Doub.

In the **Middle–Late Carboniferous**, the longitudinal-parallel disposition of ribs is inherited by gymnosperms, being typical of new genera and species of quasi-bisaccate and saccate pteridosperm miospores *Kosankeisporites* (Bharad.) Peppers, *Complexisporites* Jizba, *Lueckisporites* (Pot. et Kl.) Kl. (Fig. 1, fig. *14*), *Striatopodocarpites* (Sed.) Hart, *Striatoabieites* (Sed.) Hart (Fig. 1, fig. *13*), *Strotersporites* Wils., and nonsaccate *Aumancisporites* (Alp.) Jans and *Vittatina* (Lub.) Jans (Fig. 1, fig. *15*). Similar costate ornamentation is also characteristic of monosaccate pollen of the genus *Striomonosaccites* Bharad. (Fig. 1, fig. *16*).

A brief review of striation patterns in Carboniferous miospores shows that five types of striated ornamentation were inherited from the Devonian ancestors. These are the radial, chaotic, spiral, triangular-parallel, and longitudinal-parallel striation characteristic of pteridosperm miospores. New glomerate and concentric striation types appeared in pteridosperms only, whereas the longitudinal-parallel type of sculpturing of pteridosperm spores became widespread later in the Permian gymnosperms.

The **Permian** palynofloras of the Euramerian region, Angaraland and Gondwanan Kingdom are dominated by miospores with well-developed striation (72 form genera).

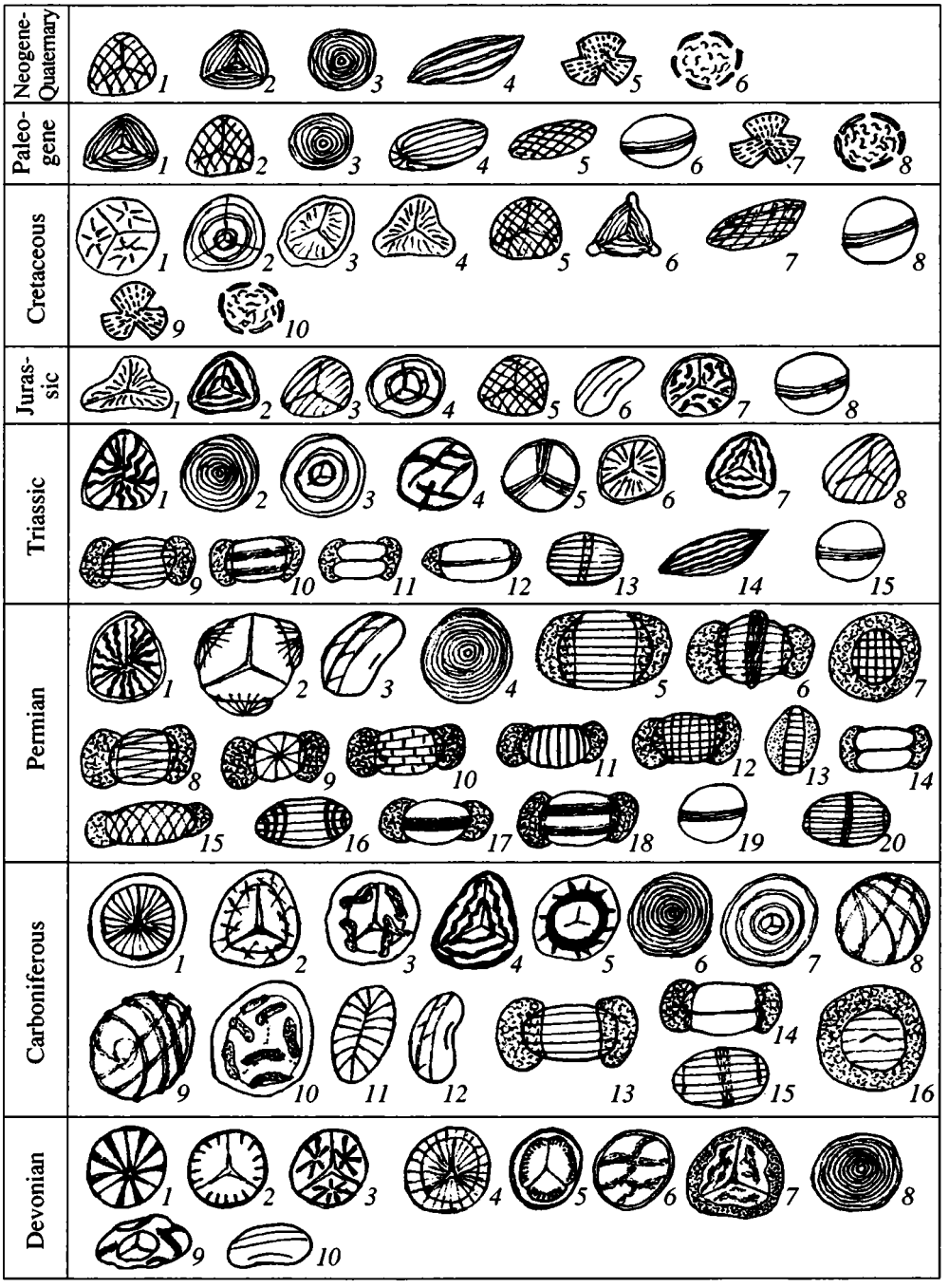
Among Permian sporogenous plants, striation is mainly characteristic of Gondwanan pteridosperms. These are *Circulisporites parvus* (de Jersey) Norris with the spiral striation (Fig. 1, fig. *4*), monolete *Kendosporites striatus* (Salayja) Surang et Chanda with the longitudinal-parallel costate ornamentation, and *Tigrisporites playfordi* de Jersey et Hamilton (Upper Permian of western Pakistan, Fig. 1, fig. *1*) with the radial striation. Permian miospores *Dulhuntyspora dulhunti* Pot. from Australia and Africa are characterized by intraradial striation (Fig. 1, fig. *2*). Among pteridosperms of the Euramerian region, striated sculpturing was characteristic of monolete spores *Columinisporites heyleri* (Doub.), which have longitudinal-parallel vermiculate costae connected by diagonal bridges (Fig. 1, fig. *3*).

Pteridosperms of the Euramerian and Angaraland regions had mainly longitudinal-parallel striation of the miospore body (Fig. 1, fig. *6*, Plate, figs. *6–8*). Similar exine construction is typical of many representatives of genera *Striatodiplonites* Elf., *Protohaploxipinus* (Samoil.) Hart, *Striatopodocarpites* (Sed.) Hart, *Striatoabieites* (Sed.) Hart, *Strotersporites* (Wilson) Kl., *Costapollenites* Tschudy, *Fusacolpites* Bose et Kar (Fig. 1, fig. *13*), *Marsupipollenites* Balme et Hennely, *Vittatina* (Lub.) Jans. (Fig. 1, fig. *20*), and others. The longitudinal-parallel striation of gymonosperm taxa is characterized by the parallel arrangement of morphological elements extending along the longitudinal axis of the miospore body. Narrow costae are usually numerous, suitable for discrimination of taxa with longitudinal costae wedging out either in the middle part of miospore or in its lateral parts [e.g., *Striatopodocarpites limnisaccata* (Wils.) Hart, Fig. 1, fig. *8*], with two wide longitudinal proximal corymbs separated by diametric laesura [corymbose *Lueckisporites* (Pot.) et Kl.) Kl. (Fig. 1, fig. *14*)], and forms with taeniate striation having 2–6 (*Lunatisporites* Leischik) Jans, Fig. 1, fig. *18*) or one (*Scutasporites* Klaus, Fig. 1, fig. *17*) wide taenia of a

belt shape. Sometimes, the proximal costate ornamentation is complicated by perpendicular costae, thus representing the longitudinal-perpendicular striation type as in *Hamiapollenites* (Wils.) Tschudy et Kosanke and *Distriatites* Bharad. (Fig. 1, figs. *6, 12*).

An intricate system of costae appeared in miospores of Gondwanan pteridosperms. *Rhizomaspora* Wils. is characterized by the radial striation (Fig. 1, fig. *9*). Representatives of form genera *Primuspollenites* Tiwari, *Verticipollenites* Bharad., *Lahirites* Bharad., and *Hindipollenites* Bharad. exhibit horizontal bands at the proximal face of miospore body, which are crossed by vertical septa forming the reticuloid striation (Fig. 1, fig. *10*). In contrast to longitudinal-parallel striation of a widespread occurrence, miospores of *Striapollenites* Bharad. have vertical-parallel striation (Fig. 1, fig. *11*). Miospores of the genus *Tornopollenites* Morgan bear narrow striation elements coiling around their bodies like a thread around spindle and characterize the fusiform striation type (Fig. 1, fig. *15*).

Different ornamentation types of monosaccate miospores are represented by horizontal striation at the



proximal face of *Stiromonosaccites circularis* Bharad. et Salujha and *Potonieisporites striatus* Koltch., by vertical striation of *Mobuitasporites ovatus* Bose et Kar, and by reticulate striation at distal and proximal faces of *Distriomonosaccites ovalis* Bharad. et Salujha (Fig. 1, fig. *7*). The last type is also developed in non-saccate miospores of *Duplivittatina persecta* f. *angusticostata* (Sauer) Koloda. Pole areas of non-saccate *Decussatisporites lucifer* Bharad. et Salujha and *Weylandites* Bharad. et Salujha are covered with vertical and horizontal taenae intricately interlacing (Fig. 1, fig. *16*). A very fine equatorial-concentric striation is typical of miospores *Protoclassopollis excretorius* Stanitchn. (Fig. 1, fig. *19*).

Thus, 14 types of striation are distinguishable in Permian miospores. The highest diversity (9 types) is characteristic of gymnosperms. Pteridosperm miospores inherited some types of striation that first appeared in the Devonian and Carboniferous spores.

The **Triassic** time, especially the Late Triassic, marks a decrease in abundance of striate miospore genera. The Triassic palynofloras include 45 striate form genera, which were most diverse in the Early–Middle Triassic. Sporogenous plants exemplify the recurrence of striation types known in the Devonian and Carboniferous floras. Par example, *Stereisporites radiatus* Schulz. with radial arrangement of sculptural elements at the proximal face is close in morphology to the Devonian *Emphanisporites robustus* Mc Greg. (Fig. 1, fig. *6*). In Middle–Upper Triassic red-beds and coaliferous sediments of the North European Russia, Varyukhina (1971, Plate IV, fig. *1*) found *Periplecotriletes* cf. *tortilis* Fadd. with a chaotic arrangement of muri at the spore surface, the form resembling Devonian species of this genus (Fig. 1, fig. *4*).

The Triassic palynofloras include also morphotypes first recorded in the Paleozoic. These are *Circulisporites parvus* (de Jersey) Norris (Fig. 1, fig. *2*) with spiral striation, *Dulhuntyispora* Pot. with the intraradial striation, *Concentricisporites nevesii* Antonescu (Fig. 1, fig. *3*) and *Polycingulatisporites crenulatus* Playf. et Dett. with concentric-circular ornamentation of the spore surface. *Onychium amplectiformis* K.-M. (Fig. 1, fig. *7*) from the Middle–Upper Triassic sediments has triangular-parallel arrangement of vermiculate muri and resembles the Carboniferous species *Cal-*

Fig. 1. Main morphotypes of striate miospore from different geological periods of the Middle–Late Phanerozoic (compiled by N.N. Podgainaya).

Devonian: (*1*) *Emphanisporites micrornatus* Richard. et Lister, 1969; (*2*) *E. protophanus* Richard. et Ioanides, 1986; (*3*) *E. schultzii* (Schultz) Mc Greg., 1973; (*4*) *Hymenozonotriletes radiatus* Naum., 1953; (*5*) *Retusotriletes striatus* (Peppers) Pepp. et Damb., 1969; (*6*) *Periplecotriletes tortus* Egorova, 1974; (*7*) *Clivosispora verrucata* var. *convoluta* Mc Greg. et Camf., 1977; (*8*) *Chomotriletes vedugensis* Naum., 1953; (*9*) *Archaeozonotriletes polymorphus* Naum., 1953; (*10*) *Limbomonoletes lenis* Arch., 1985.

Carboniferous: (*1*) *Archaeozonotriletes intrastriatus* Playf., 1971; (*2*) *Velamisporites irrugatus* Playf., 1978; (*3*) *Cincturasporites* (Hacq. et Barss) Byvsch., 1985; (*4*) *Callisporites* Butt. et Will., 1958; (*5*) *Radiizonates striatus* (Knox) Stapl. et Jans., 1964; (*6*) *Chomotriletes multivittatus* Playf., 1978; (*7*) *Concentricisporites concentricus* Byvsch., 1976; (*8*) *Vestispora lucida* (Butt. et Will.) Pot., 1960; (*9*) *Proprisporites laevigatus* Neves, 1958; (*10*) *Arcuatisporites* Tet., 1984; (*11*) *Radialetes costatus* Playf., 1963; (*12*) *Columinisporites heyleri* Doub., 1973; (*13*) *Striatoabieites* (Sed.) Hart, 1964; (*14*) *Lueckisporites* (Pot. et Kl.) Kl., 1963; (*15*) *Vittatina* (Lub.) Jans., 1962; (*16*) *Striomonosaccites* Bharad., 1962.

Permian: (*1*) *Tigrisporites playfordi* de Jersey end Hamilton, 1967; (*2*) *Dulhuntyispora dulhuntyi* Pot., 1956; (*3*) *Columinisporites heyleri* Doub., 1973; (*4*) *Circulisporites parvus* (de Jersey) Norris, 1965; (*5*) *Protohaploxipinus* (Samoil.) Hart, 1964; (*6*) *Hamiapollenites* (Wils.) Tschudy et Kosanke, 1966; (*7*) *Distriomonosaccites ovalis* Bharad. et Salujha, 1964; (*8*) *Striatopodocarpites limnisaccata* (Wils.) Hart, 1964; (*9*) *Rhizomaspora* Wils., 1962; (*10*) *Verticipollenites* Bharad., 1962; (*11*) *Striapollenites* Bharad., 1962; (*12*) *Distriatites* Bharad., 1962; (*13*) *Fusacolpites* Bose et Kar, 1966; (*14*) *Lueckisporites* (Pot. et Kl.) Kl., 1963: (*15*) *Tornopollenites* Morgan, 1971; (*16*) *Decusatisporites lucifer* Bharad. et Salujha, 1964; (*17*) *Scutasporites* Kl., 1963; (*18*) *Lunatisporites* (Leschik) Jans., 1962; (*19*) *Protoclassopollis* Stanitch., 1985; (*20*) *Vittatina* (Lub.) Jans., 1962.

Triassic: (*1*) *Tigrisporites playfordi* de Jersey end Hamilton, 1967; (*2*) *Circulisporites parvus* (de Jersey) Norris, 1965; (*3*) *Concentricisporites nevesii* Antonescu, 1970; (*4*) *Periplecotriletes* cf. *tortilis* Fadd., 1965; (*5*) *Naumovaspora striata* Jans., 1962; (*6*) *Stereisporites radiatus* Schulz, 1970; (*7*) *Onychium amplectiformis* K.-M., 1956; (*8*) *Contignisporites dunrobinensis* (Couper) Schulz, 1967; (*9*) *Striatoabieites* (Sed.) Hart, 1964; (*10*) *Lunatisporites* (Leschik) Jans., 1962; (*11*) *Lueckisporites* (Pot. et Kl.) Kl., 1963; (*12*) *Chordasporites* Kl., 1960; (*13*) *Vittatina* (Lub.) Jans, 1962; (*14*) *Gnetaceaepollenites* Thirgart, 1958; (*15*) *Classopollis* (Pflug) Pocock et Jans., 1961.

Jurassic: (*1*) *Triparlina variabilis* Mal., 1949; (2) *Duplexisporites anagrammensis* (K.-M.) Sem., 1970; (*3*) *Contignisporites dunrobinensis* (Couper) Schulz, 1967; (*4*) *Polycingulatisporites* Simons et Kedves, 1965; (*5*) *Cicatricosisporites brevilaesuratus* Couper, 1958; (*6*) *Azonomonoleles* Vishnu-Mittre, 1954; (*7*) *Rugulatisporites clivosus* (Rolch.) Sem., 1970; (*8*) *Classopollis classoides* Pflug, 1953.

Cretaceous: (*1*) *Stenozonotriletes radiatus* Chlon., 1964; (*2*) *Taurocusporites reduncus* (Bolch.) Stover, 1962; (*3*) *Selaginella sibiriciformis* Griaz., 1965, (*4*) *Tripartina variabilis* Mal., 1949; (*5*) *Cicatricosisporites australiensis* (Cook.) Pot., 1958; (*6*) *Anemia* Swartz, 1806; (*7*) *Gnetaceaepollenites* Thiergari, 1938; (*8*) *Classopollis classoides* Pflug, 1953; (*9*) *Acer tener* Samoil., 1965; (*10*) *Ulmoideipites krempii* Anderson, 1960.

Paleogene: (*1*) *Anemia* Swartz, 1806; (2) *Cicatricosisporites* Pot. et Golletich, 1933; (*3*) *Glomeratia circula* (Wolf) Lubomir., 1977; (*4*) *Schizaeae* J.E. Smith, 1798; (*5*) *Gnetaceaepollenites* Thiergart, 1938; (*6*) *Classopollis classoides* Pflug, 1953; (*7*) *Acer tener* Samoil., 1965; (*8*) *Ulmoideipites krempii* Anderson, 1960.

Neogene–Quaternary: (*1*) *Cicatricosisporites* Pot. et Golletich, 1933; (*2*) *Anemia* Swartz, 1806; (*3*) *Glomeratia circulla* (Wolf) Lubomir., 1977; (*4*) *Ephedra fusiformis* Shakhm., 1965; (*5*) *Acer* sp.; (*6*) *Ulmus rotundata* Anan., 1974.

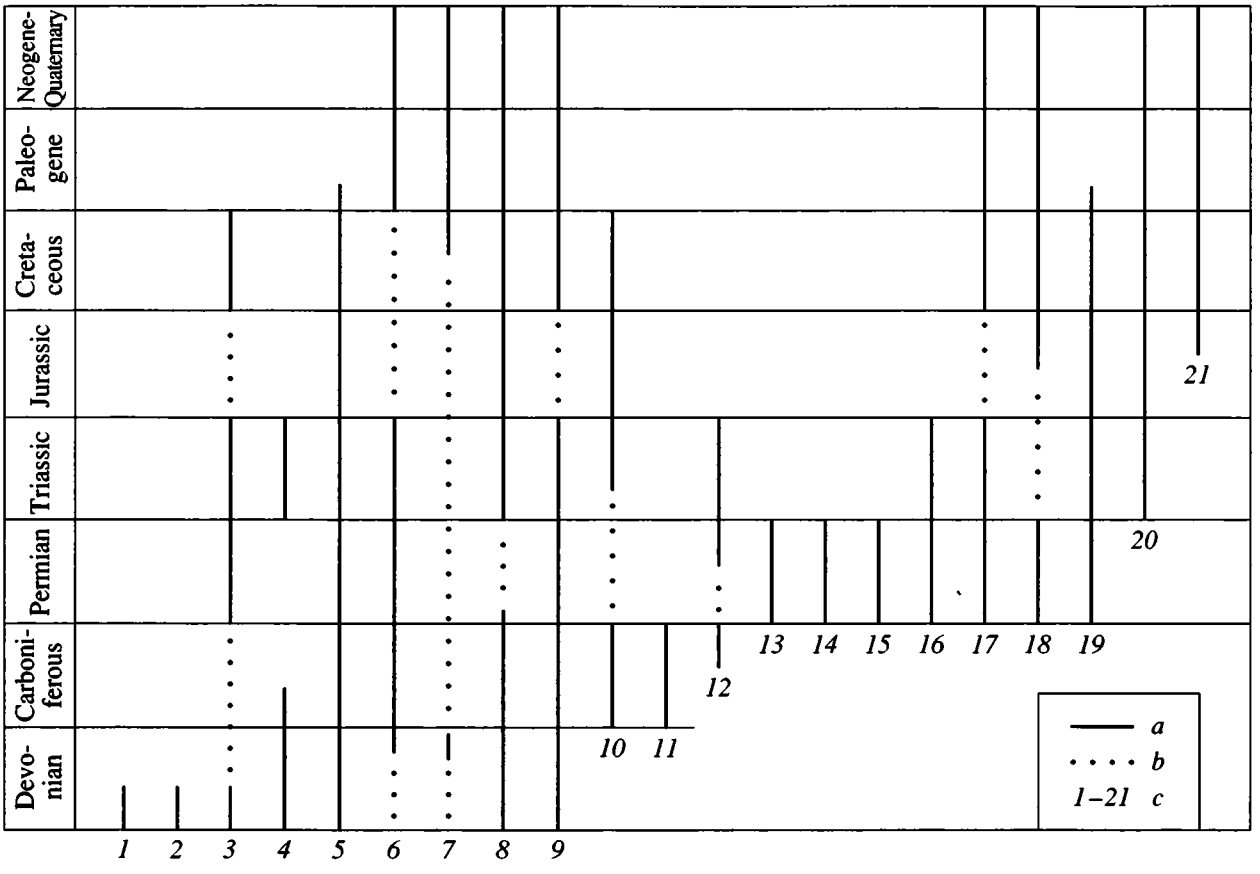


Fig. 2. Stratigraphic distribution of main striation types (compiled by N.N. Podgainaya): (*a*) recorded, (*b*) not recorded; (*c*) striation types (*1*) radial-extending, (*2*) radial-equatorial, (*3*) intraradial, (*4*) chaotic, (*5*) radial, (*6*) spiral, (*7*) streaming, (*8*) triangular-parallel, (*9*) longitudinal–parallel, (*10*) concentric–circular, (*11*) glomerate, (*12*) corymbose, (*13*) striate-reticulate, (*14*) vertical-parallel, (*15*) longitudinal-wedging, (*16*) taeniate, (*17*) fusiform, (*18*) distal–proximal, (*19*) equatorial-concentric, (*20*) diagonal, (*21*) cerebriform.

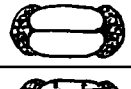
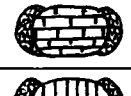
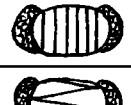
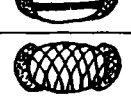
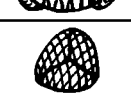
lisporites nux Butt. et. Will. (Fig. 1, fig. *4*, Carboniferous). Species *Tigrisporites playfordi* de Jersey and Hamilton with the radial striation, which was first found in the Upper Permian sediments of Pakistan, is typical of the Middle–Upper Triassic in Europe.

Many form genera of Triassic gymnosperms had miospores covered by longitudinal-parallel striae inherited from the Permian Epoch. These were *Striatites* (Pant) Bharad., *Striatoabieites* (Sed.) Hart, *Stroter-sporites* (Wils.) Kl., *Protohaploxipinus* (Samoil.) Hart, *Lueckisporites* (Pot. et. Kl.) Kl., *Lunatisporites* (Leschik) Jans., and *Vittatina* (Lub.) Jans. A significant component of Triassic palynofloras is the genus *Gneta-ceaepollenites* Thiergart represented by fusiform miospores with vermiculate longitudinal-parallel costae converging toward poles (Fig. 1, fig. *14*). The Late Triassic palynofloras include typical *Classopollis* (Pflug) Pocock et Jans with equatorial–concentric striation.

The above data show that Triassic striated morpho-types are most inherited from the Paleozoic. Only two new types of striation originated during the Triassic. These are the fine radial-parallel costate ornamentation confined to radial regions in *Naumovaspora striata* Jans. and diverging in a form of three bunches from the spore pole toward equator (Fig. 1, fig. *5*) and the diagonal striation of rounded subtriangular miospores *Con-tignisporites dunrobinensis* (Couper) Schulz having 7 to 9 muri oriented parallel to one of the spore faces (Fig. 1, fig. *8*).

The **Jurassic** palynofloras include 22 form genera with species ornamentation of different types, characteristic of which is again the recurrence of morpho-types. The Early Jurassic is marked by development of typical and stratigraphically significant *Tripartina vari-abilis* with the radial striation (Fig. 1, fig. *1*). The Jurassic species *Duplexisporites anagrammensis* K.-M. with triangular-parallel arrangement of vermiculate costae at the proximal face (Fig. 1, fig. *2*, Jurassic) resembles the Carboniferous *Callisporites nux* Butt. et. Will. (Fig. 1, fig. *4*, Carboniferous) and Triassic *Onychium amplecti-formis* K.-M. (Fig. 1, fig. *7*, Triassic), which still occur in the Lower Jurassic sediments. Similar ornamentation type is also characteristic of *Pteris multinoides* Bolch.

Appendix to Fig. 2. Main striation types of Upper Paleozoic, Mesozoic, and Cenozoic miospores (compiled by N.N. Podgainaya).

Ordred number	Striation type	Type species	Schematic sketch
1	Radial-extending	*Emphanisporites micrornatus* Richardson et Lister, 1969	
2	Radial-equatorial	*Emphanisporites protophanus* Richardson et Ioanides, 1986	
3	Intraradial	*Emphanisporites schultzii* (Schultz) Mc Gregor, 1973; *Stenozonotriletes radiatus* Chlonova, 1964	
4	Chaotic	*Periplecotriletes tortus* Egorova, 1974	
5	Radial	*Hymenozonotriletes radiatus* Naumova, 1953; *Archaeozonotriletes intrastriatus* Playford, 1971	
6	Spiral	*Chomotriletes vedugensis* Naumova, 1953	
7	Streaming	*Acer tener* Samoilovich, 1965	
8	Triangular-parallel	Callisporites Butterworth et Williams, 1958; Mohria Swartz, 1806	
9	Longitudinal-parallel	*Limbomonoletes lenis* Archangelskaja, 1985; Vittatina (Luber) Jansonius, 1962	
10	Concentric-circular	*Concentricisportes concentricus* Byvscheva, 1976	
11	Glomerate	*Vestispora lucida* (Butterworth et Williams) Potonie, 1960	
12	Corymbose	*Lueckisporites virkkiae* (Potonie et Klaus) Klaus, 1963	
13	Striate-reticulate	Verticipollenites Bharadwaj, 1962	
14	Vertical-parallel	Striapollenites Bharadwaj, 1962	
15	Longitudinal-wedging	*Striatopodocarpites limnisaccata* (Wilson) Hart, 1964	
16	Taeniate	Lunatisporites (Leschik) Jansonius, 1962	
17	Fusiform	Tornopollenites Morgan, 1971	
18	Distal-proximal	*Distriomonosaccites ovalis* Bharadwaj et Salujha, 1964; *Cicatricosisporites brevilaesuratus* Couper, 1958	
19	Equatorial-concentric	*Classopollis classoides* Pflug, 1953	
20	Diagonal	*Contignisporites dunrobinensis* (Couper) Schulz, 1967	
21	Cerebriform	*Ulmus rotundatus* Ananova, 1974	

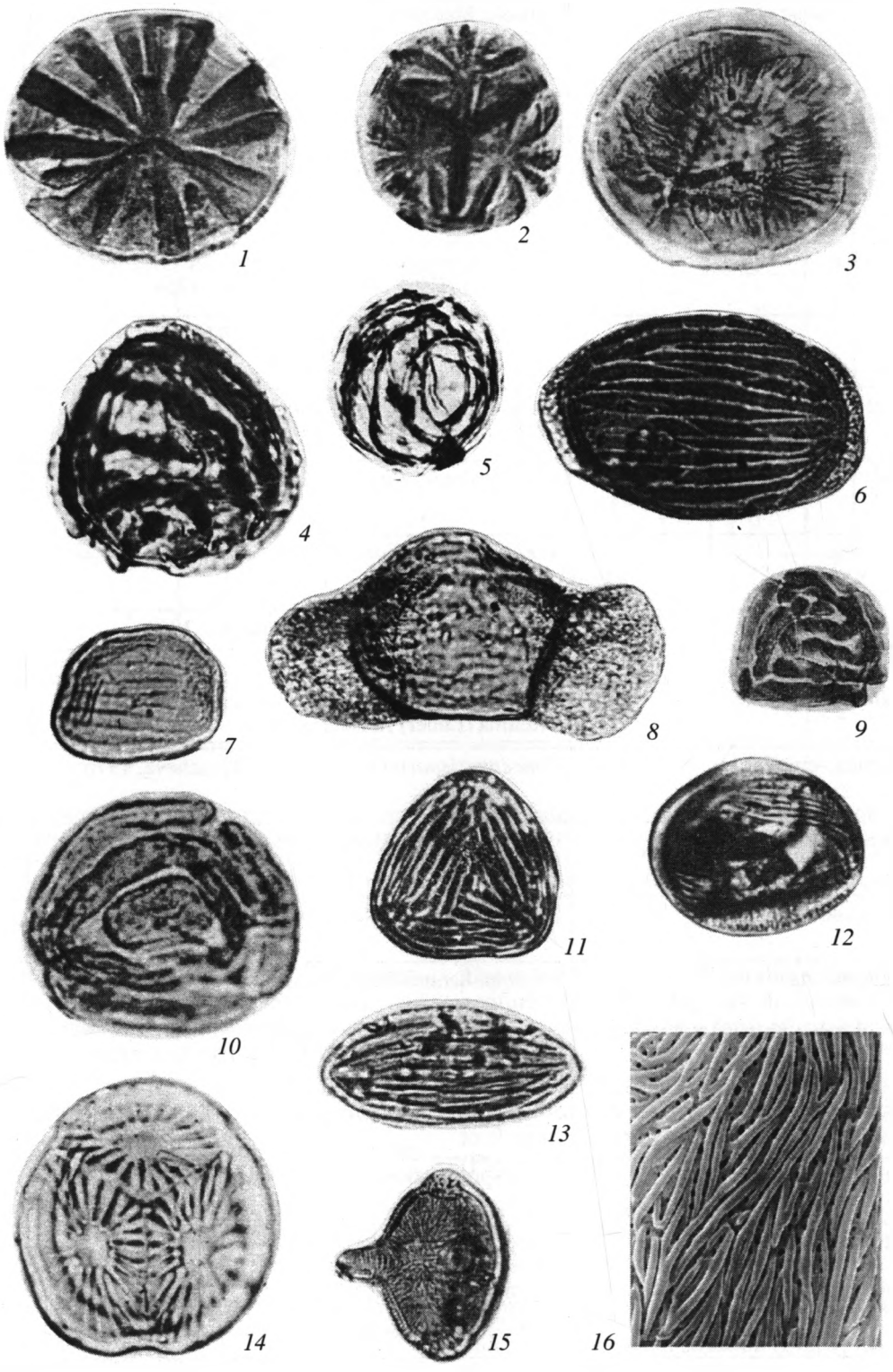


Plate. Phanerozoic striate miospore and striation types (compiled by N.N. Podgainaya).

(*1*) *Emphanisporites robustus* Mc Greg., radial-extending (Arkhangel'skaya, 1980, Lower Devonian, Plate VIII, fig. 1, ×750); (*2*) *Emphanisporites schultzii* Mc Greg., intraradial (Arkhangel'skaya, 1980, Lower Devonian, Plate VIII, fig. 3, ×750); (*3*) *Archaeozonotriletes intrastriatus* Playf., radial (Playford, 1971, Lower Carboniferous, Plate 15, fig. 12, ×500); (*4*) *Proprisporites laevigatus* Neves, glomerate (Teteryuk, 1984, Lower–Middle Carboniferous, Plate 87, fig. 5, ×500); (*5*) *Vestipora profunda* Wols. et Hoff., glomerate (collection by E. G. Shvartsman, Middle Carboniferous, ×500; (*6*) *Protohaplopinites vulgaris* Efr., longitudinal-wedging (collection by N. N. Podgainaya, Lower Permian, ×500); (*7*) *Vittatina striata* Lub., longitudinal-parallel (collection by

The distal face of the last miospore species can also be ornamented by costae arranged parallel to any face of the rounded–triangular spore (Plate, fig. *9*) that demonstrates therefore two striation types.

Species with the radial striation, which cross the Triassic–Jurassic boundary, are *Zebrasporites fimbriatus* Kl., *Tigrisporites playfordi* de Jersey and Hamilton, and *Leiotriletes eximinus* Bollch.

The diagonal striae arrangement is observable at distal faces of miospores *Contignisporites dunrobinensis* (Couper) Schulz (Fig. 1, fig. *3*) and *Camptotriletes anagrammensis* K.-M. Concentric-circular striation is characteristic of species *Polycingulatisporites circulus* Simoncs et Kedves (Fig. 1, fig. *4*), *Cingulatisporites foveolatus* Coup., and *Stereisporites trizonatus* Schulz. The longitudinal-parallel striation is recognized in bilateral spores of the genus *Azonomonoletes* Vishnu-Mittre (Fig. 1, fig. *6*), which have broad, sometimes dichotomizing bands. The distal–proximal mutually perpendicular ribs characterize striation of *Cicatricosisporites brevilaesuratus* Couper (Fig. 1, fig. *5*). In their majority, the Jurassic gymnosperms are lacking striate miospore. A fine equatorial-concentric striation is characteristic of *Classopollis classoides* Pflug. only, which is a typical taxon since the Late Jurassic.

Thus, seven types of striation are inherited by Jurassic palynoflora from previous epochs. During this period, a new cerebriform striation appeared in *Rugulatisporites clivosus* (Bolch.) Sem. (Fig. 1, fig. *7*). In this species, radial sectors of the proximal face are ornamented with narrow vermiculate striae. Subsequently, this striation type becomes typical of Cretaceous elm-like pollen (Fig. 1, fig. *10*, Cretaceous).

The **Cretaceous** striate miospores are characteristic of 37 form genera, most of which belong, as in the Carboniferous, to pteridosperms.

Spores *Tripartina variabilis* Mal. (Fig. 1, fig. *4*) with radial striation cross the Jurassic–Cretaceous boundary. A noticeable component of palynofloras is the recurrent morphotype *Stenozonotriletes radiatus* Chlon. with the intraradial arrangement of striae (Fig. 1, fig. *1*, Cretaceous, Plate, fig. *14*), the taxon resembling the Devonian species *Emphanisporites schulzi* Mc Greg. (Plate, fig. *2*). Rugulae radially arranged at the proximal face are typical of *Selaginella sibiriciformis* Griaz. (Fig. 1, fig. *3*), *Cooksonites medullatus* Chlon., and *Rouseisporites triangularis* Pocock. The concentric-circular striation is observable in *Taurocusporites reduncus* (Bolch.) Stover (Fig. 1, fig. *2*) and distal–proximal cicatricose ornamentation is typical of *Cicatrisisporites australiensis* (Cook) Pot. (Fig. 1, fig. *5*). The triangular-parallel costate sculpturing is characteristic of form taxa *Appendicisporites matesovae* (Bolch.) Norris, *Pelletieria striata* Bolch. (Plate, fig. *11*), *Rufflordia goepperti* Seward, and *Anemia* Swartz. (Fig. 1, fig. *6*). The diverse pollen taxa with longitudinal-parallel striation are *Welwitschiapites magniolobatus* Bolch., *Corniculatisporites striatus* (Kuv., *Chlonovaia sibirica* Elsk, and *Ephedripites* sp. The fusiform striation is typical of *Gnetaceaepollenites laevigataeformis* (Bolch.) Verb. (Fig. 1, fig. *7*) and *Steevesipollenites virgatus* Schwetzova.

Pollen of angiosperm species that appeared in the terminal Early Cretaceous frequently occurs in association with abundant pollen of *Classopollis classoides* Pflug (Fig. 1, fig. *8*, Plate, fig. *12*) that retains the equatorial-concentric striation. Pollen of many pioneering angiosperm species has protective striate structures. Most commonly, the latter are represented by fine radial striation that first originated in Devonian sporogenous plants and then disappeared for a long period. Striation of this kind is characteristic of *Tricolporites striatellus* Mtsch., *Borealipollenites bratzevae* Chlon., *Cranwellia striata* (Coup.) Srivastava, *Elytranthe striatus* Coup., *Parviprojectus striata* Mtsch. (Plate, fig. *15*), and *Acer tener* Samoil. (Fig. 1, fig. *9*). Plants producing elm-type pollen *Ulmoideipites krempi* Anderson (Fig. 1, fig. *10*) were likely well adopted to arid environments, because their pollen grains are covered by short narrow cerebriform-vermiculate muri.

Thus, there were 10 types of sculpturing elements arrangement in Cretaceous striate miospores. Most of these types, except for diagonal and cerebriform striation that appeared in the Mesozoic (Fig. 2), represent recurrence of Paleozoic striation. Streaming ornamentation type first recorded in Devonian spores reappeared in Cretaceous angiosperms after a long period of latent state.

At the beginning of the *Paleocene*, quantity of "striated genera" decreased. In total, there are known 20 genera of Paleogene striate miospores, ten of which are inherited from the Cretaceous.

The form taxa of pteridosperms that crossed the Cretaceous–Paleogene boundary are *Anemia* Swartz with triangular-parallel striation, *Cicatricosisporites* Pot. with distal–proximal striation, and Schizeae with the longitudinal-parallel or fusiform costate ornamen-

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N. N. Podgainaya, Lower Permian, ×500); (*8*) *Striatoabieites* sp., longitudinal-parallel (collection by N. N. Podgainaya, Lower Permian, ×500); (*9*) *Duplexisporites anagrammensis* (K.-M. et Bolch.) Shug., triangular-parallel (Timoshina and Men'shikova, 1980, Lower Jurassic, Plate IV, fig. 11, distal face, ×600); (*10*) *Polycingulatisporites* sp., concentric-circular (Yaroshenko, 1978; Middle Triassic, Plate VI, fig. 3, ×1000); (*11*) *Pelletieria striata* Bolch., triangular-parallel (Bitytskaya *et al.*, 1973, Lower Cretaceous, Plate VII, fig. 3, ×600); (*12*) *Classopollis* sp., equatorial-concentric (Komarova, 1973, Upper Cretaceous, Plate II, fig. 16, ×900); (*13*) *Ephedra* cf. *tenera* Shakhm., longitudinal-parallel (Panova *et al.*, 1990, upper Oligocene, Plate LXIV, fig. 7, ×600); (*14*) *Stenozonotriletes radiatus* Chlon., intraradial ((Khlonova, 1980, Cretaceous, Plate 2, fig. 8, ×900; (*15*) *Parviprojectus striatus* Mtch., radial-striate (Mchedliashvili, 1965, Upper Cretaceous, Plate 1, fig. 19, ×500; (*16*) *Dryas integrifolia* Vakl. Subsp. *crenulata* (Tuz.) Kozhevn., streaming (collection by D.B. Arkhangel'skiĭ, Yakutia, Evata Rock, Recent, pollen surface, ×10000.

tation (Fig. 1, figs. *1, 2, 4*). Genera *Gnetaceaepollenites* Thierg. with fusiform striation, *Ephedripites* Bolch., *Weiwitschiapites* Bolch. with longitudinal-parallel striation, and *Classopollis* Pflug. with the equatorial-concentric costate sculpturing occur among gymnosperms. The radial-striate (*Elytranthe striatus* Coup., *Tricolporites striatellus* Mtsch., *Acer tener* Samoil,) or longitudinal-striate ornamentations (*Tricolporopollenites villensis* (Th.) Pflug, *Quercus sparsa* Mart., *Rhoipites striatus* Ked.) prevail in angiosperm pollen (Fig. 1, fig. *7*). Pollen of the genus *Ulmus* L. retains the cerebriform striation type (Fig. 1, fig. *8*). Similar striation is also characteristic of the genus *Zelkova* L. The Oligocene sediments of West Siberia contain miospores *Glomeratia circula* (Wolf.) Lubomir. with fine concentric striation patterns at the surface of inaperturate spherical grains (Fig. 1, fig. *3*). In density of concentric elements, this striation is similar to that of *Chomotriletes vedugensis* Naum., *Ch. multivittatus* Playf., and *Circulisporites parvus* (de Jersey) Norris from the Upper Paleozoic and Triassic sediments. Accordingly, it can be referred to the spiral type (Fig. 1, fig. *8*, Devonian, fig. *6*, Carboniferous, fig. *4*, Permian, fig. *2*, Triassic).

The genus *Classopollis* Pflug with equatorial-concentric striation became extinct in the Paleocene. Other striation types remain characteristic of the Neogene–Quaternary and recent miospore species (Fig. 2).

The *Neogene–Quaternary* floras demonstrate steadily increasing abundance of angiosperm genera with striated pollen exine. They include 104 striated form genera.

The **Miocene** palynological assemblages include spores *Cicaricosisporites* Pot. et Gelletich with the distal-proximal type of costate ornamentation (Fig. 1, fig. *1*) inherited from the Jurassic, *Corugatisporites semivallatus* W. Kr. and *Anagramma leptophylla* (L.) Link., both having triangular-parallel costate sculpturing characteristic of many species of Phanerozoic pteridosperms. The triangular-parallel and distal–proximal types of costate ornamentation are typical of many recent species of genera *Anemia* Swartz and *Mohria* Swartz dwelling in near-equatorial regions. The fusiform striation occurs in monolete spores of genera *Schizaea* Smith (*S. digitata* (L.) Sw., *S. laegigata* Mett.) populating the same regions. In representatives of *Magnolia* L., grana of the fine-grained exine are fused into vermiculate structures, whereas closely arranged costae of *Menyanthus trifoliate* L. pollen look like twisted. The fossil and recent palynofloras include also pollen grains with spiral arrangement of costae. Pollen grains of *Glomeratia circula* (Wolf) Lubomir. and *Crocus korolkowi* Maw. et Regel occurring respectively in the Pliocene and Recent palynofloras of Germany are striated in this manner. The cerebriform ornamentation is characteristic of the Ulmaceae pollen (Fig. 1, fig. *6*).

The longitudinal–parallel striation peculiar of the Permian *Vittalina* species is inherited by representatives of two recent gymnosperm genera *Ephedra* and *Welwitschia* (Fig. 1, fig. *4*, Plate, fig. *13*), and by recent angiosperm monocotyledons *Spathiphyllum, Amorphophallus, Pistia*, and *Arisarum*.

Under optical microscope, streaming ornamentation is distinguishable in most of striate pollen grains of angiosperms. Under greater magnification of about 10000, it is possible to distinguish muri at the exine surface of these grains, which are fused, interlaced, dichotomizing, and perforated sometimes (Plate, fig. *16*). In the Neogene–Quaternary palynofloras, there are 70 of 104 striate genera with the diverse streaming ornamentation. For example, *Polemonium acutiflorum* Willd has the reticulate–streaming ornamentation with short or long, sometimes curved subcircular striae connected by thin anastomosing bridges consisting of minute rods. *Lucium barbatum* L. bears short meridional striae composed of closely spaced rods, while *Spiraea crenata* L. is characterized by the oblique-striate ornamentation composed of long striae hardly visible in immersion liquid. Exine of *Helianthemum canum* (L.) Baumg is covered by small tubercles arranged in chains, and vermiculate-striate sculpture of *Listera cordata* (L.) R. Br. is composed of vermiculate-parallel muri uniformly oriented around lumina. Thus, all striation types recorded in the Neogene–Quaternary and recent palynofloras are inherited from past epochs (Fig. 2).

CONCLUSION

The analyzed mophogenesis and stratigraphic distribution of Phanerozoic miospores with different striation types leads to the following conclusions.

Most remarkable evolutionary changes in morphology of striate miospores are detectable from arrangement of striate sculptural elements and their orientation relative to symmetry axes and equator.

First miospores with simplest (radial and radial-extending) striation subsequently gave way to miospores with more intricate arrangement of various sculptural elements, to fusion of exine projections, and to appearance of intraradial, chaotic, triangular-parallel, striate, glomerate, concentric-circular, equatorial-concentric, diagonal, distal–proximal, striate–reticulate, cerebriform, and other types of striation. Some spores exhibit combinations of two striation types.

The evolution of striation phenomenon that lasted more than 400 m.y. resulted in formation of 21 types of striate ornamentation (Fig. 2, Appendix to Fig. 2). The observation is consistent with earlier inferences (Meyen, 1987) that miospores of higher plants exhibit a limited number of striation modes, and recurrence of similar types in different eutaxa is unavoidable therefore.

The Devonian and Permian were periods of intense development of new striation types. During other geological periods, miospores inherited older sculptural elements and formation rate of new morphotypes was

low: only two new striation types originated in the Triassic, while in the Jurassic and Cretaceous epochs produced one new type each. Nevertheless, every geological period represented a certain stage in development of the striation phenomenon with its own specific distribution of various striation types.

The plant evolution was accompanied by recurrence of striation types. The recurrence reflects, as is suggested, the "...evolutionary parallelism and transitive polymorphism manifested in all morphological features of miospores from their general organization to exine details" (Meyen, 1987, p. 282).

It is established that some former striation types are inherited by miospores of new taxonomic plant groups. A most impressive example is the appearance of striate ornamentation in Mesozoic–Cenozoic miospores of angiosperms, because it was originally typical of Devonian sporogenous plants and then remained in the latent state over 200 m.y., i.e., during four geological periods (Fig. 2). This is consistent with inferences of Ch. Darwin, J. Agassiz, E. Haeckel, I. Müller, N. Vavilov, L. Berg, A. Lyubishchev, and other evolutionists that hereditary features can remain in the latent state unrestrictedly long time during evolution (Onoprienko, 1996, p. 62). The nature likely repeats most successful features and uses them in similar environments for protection of plant reproductive cells.

The analyzed development of striation and the recurrence phenomenon suggest that evolution followed the way of iterative (repeated) morphogenesis. The persistence capability was inherent not only in miospore morphotypes with particular striation (triangular-parallel, longitudinal-parallel, striate, and others), but also in characteristic structural elements of the exine sculpture (taeniae, costae, muri, striae).

The strict geochronological succession in origin and extinction of miospores with different striation types, which are confined to certain stratigraphic intervals, opens a possibility to define ranges of variations for particular sculptural elements of exine, to estimate the extent of their evolutionary plasticity, and to use these observation as an additional tool in palynostratigraphic studies. The global distribution of striate miospore taxa and elucidated stratigraphic ranges of different striation types facilitate the remote correlations.

In conclusion, I should emphasize the effectiveness of historical–morphological investigations in palynology, their theoretical and practical significance, and the necessity to study further the striation phenomenon under optical and electron microscopes. The study is undoubtedly promising to reveal new, yet unknown trends in evolution and new structural peculiarities of striate miospores.

Reviewers A.V. Goman'kov and O.P. Yaroshenko

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Devonian Paleogeography and Sedimentation Settings in the South Siberian Platform

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Abstract—The Early Paleozoic regression resulted in emergence of the Angaraland. Devonian deposits accumulated in various sedimentary basins, e.g., in gulfs far incised into the Angaraland, as well as in lagoons and lakes. In marginal areas of the platform, the Rybnaya and Pribaikal'skii sedimentary basins in the foredeep accumulated the molasse sequences during their early evolutionary stages. In the Angaraland, there were plains and uplands of different kinds, which experienced denudation. Erosion of these landforms affected various formations, the diamond-bearing kimberlite bodies included.

Key words: Siberian platform, Devonian, paleogeography, stratigraphy, lithology, Angaraland, sedimentation basin.

INTRODUCTION

The Siberian platform is one of the largest geological structures in northern Asia. In the geological section of the platform sedimentary cover, Devonian deposits are of particular interest as they occur solely in the rift or rift-related structures of the platform.

Devonian deposits on the platform were first discovered by V. Zlatkovskiĭ in 1886, and detailed information on this point was reported at the Soviet–Polish Symposium *Centennial Anniversary of Studies in Eastern Siberia and Lake Baikal Carried out by Polish Researchers* (Akulov, 1993). Later on, Devonian deposits were studied by S.V. Obruchev, A.V. Lesgaft, A.I. Anatol'eva, O.V. Vaag, M.A. Zharkov, I.I. Kolodeznikov, R.G. Matukhin, V. Vl. Menner, V.T. Mordovskiĭ, P.N. Sokolov, I.N. Sulimov, B.A. Fuks, N.M. Chumakov, É.N. Yanov, and others.

In 1969, the *Atlas of Lithological–Paleogeographic Maps*, scale 1 : 7500000, was published (*Atlas...*, 1969). The edition summarized all the information available before the mid-1960s. The work was carried out by a large group of specialists who created the base for lithological–paleogeographic study of all geological structures on the territory of the USSR.

Later, Vaag and Matukhin (1978) compiled lithological–paleogeographic schemes for the northwestern part of the Siberian platform. Matukhin (1991) generalized afterward the data on Devonian depositional environments for the whole Siberian platform. New lithological–paleogeographic maps, scale 1 : 5000000, were compiled for narrow stratigraphic ranges to illustrate the formation history for each distinguishable stratigraphic unit and corresponding deposits.

New data on the structure and occurrence mode of Devonian deposits in the southern part of the Siberian platform were obtained during the geological survey (GGS-50), scale 1 : 50000, carried out by PGO Irkutskgeologiya and Krasnoyarskgeologiya in the last existence years of the USSR. Based on reconstructions performed previously, I prepared a series of lithological–paleogeographic maps, scale 1 : 2500000, for the southern part of the platform, which are schematized in Figs. 1–4.

DEVONIAN DEPOSITS OF THE SOUTHERN ANGARALAND

The Early Paleozoic sea regression resulted in origin of the Angaraland in the Siberian platform, where it emerged in the Silurian time as a vast continental land area free of seawaters. Within the Angaraland, sedimentation brines of sea origin were replaced by infiltration waters, and this event was accompanied by erosion and denudation, the effect of which was much stronger than formerly in the whole Phanerozoic history of that structure. Ordovician and Upper Cambrian rocks were subjected to the most intensive erosion, since they occurred above the erosion base on the greater part of the territory. The analyzed thickness of these deposits in profiles based on borehole data showed that relief of the Irkutsk Amphitheater was dissected down to the depth of 300–500 m, in places to 850 m.

Throughout the Devonian, the Angaraland was surrounded by the Zaisan, Ob, Taimyr, Verkhoyansk, and Dauria seas (Fig. 1). In the southern part of the platform, Devonian deposits accumulated in sedimentary basins, which corresponded to sea gulfs incised far into the Angaraland or to lagoons and lakes. The most com-

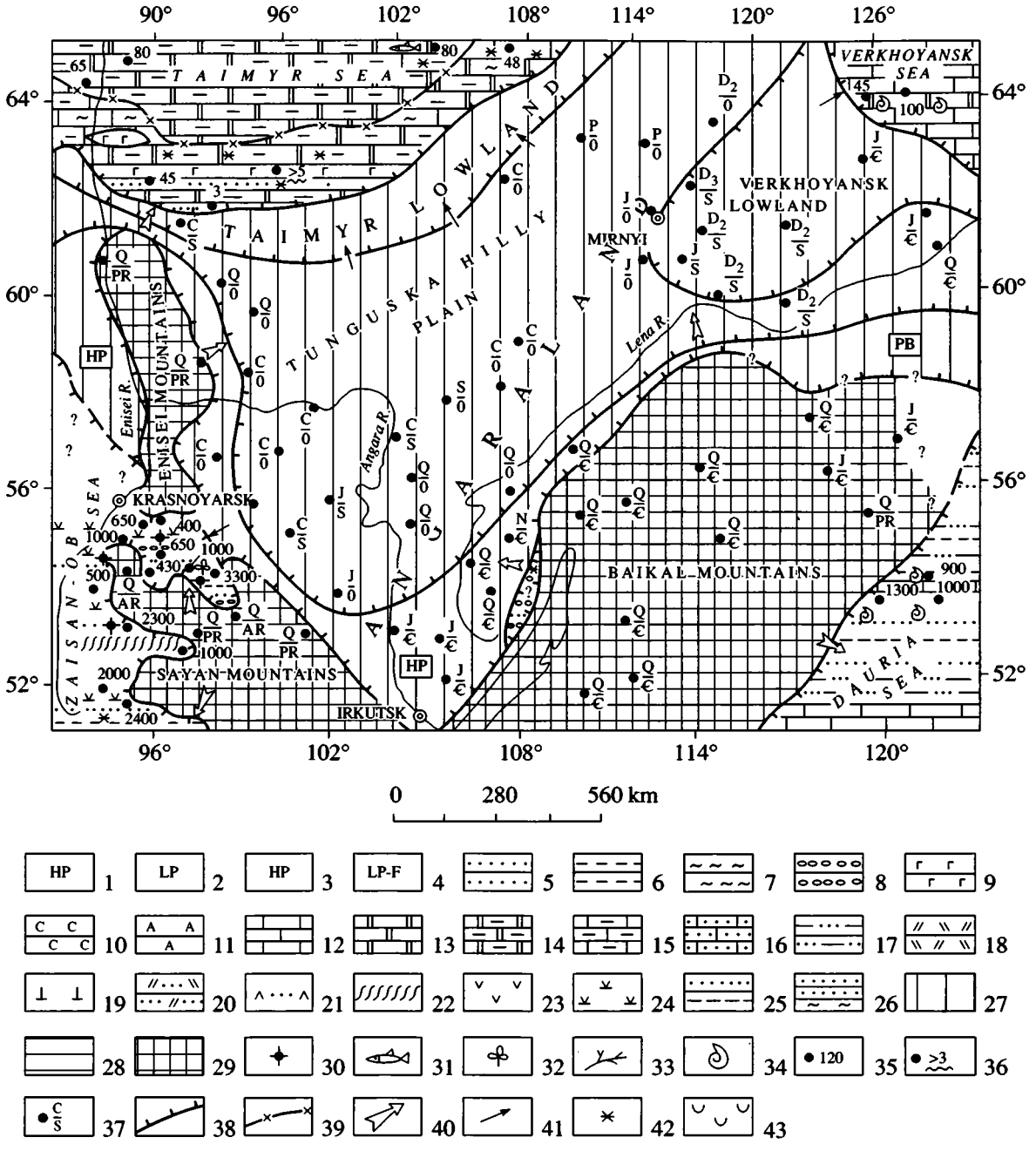

Fig. 1. Lithological–paleogeographic map of the southern Siberian platform, the Early Devonian (Lochkovian to Pragian): (1) hilly, (2) low, (3) high, and (4) low sea-flooded plains; (5) inequigranular sand; (6) siltstone; (7) clay; (8) pebble and gravel beds; (9) gypsum; (10) salt; (11) anhydrite; (12) limestone; (13) dolomite; (14) marl; (15) dolomitic limestone; (16) inequigranular calcareous sand; (17) argillaceous sand; (18) pyroclastic material; (19) basic pyroclastic material; (20) sand with pyroclastic material; (21) acid pyroclastic material; (22) intermediate pyroclastic material; (23) basic volcanogenic rocks; (24) basic lavas; (25) alternating sediments, the amount of which in the section is nearly equal; (26) alternation of sediments with considerable predominance of one type over the other; (27) Lower Paleozoic terrigenous–carbonate deposits; (28) Middle Paleozoic carbonate–terrigenous deposits; (29) Precambrian metamorphic and igneous rocks; (30) basal pipes and kimberlite fields; (31) fish fauna, (32) flora, (33) plant detritus, (34) other fauna; (35) reference borehole sections and natural exposures (thickness, m); (36) incomplete sections lacking data on their lower part; (37) rocks of given age are missing (overlapping of older sediments by younger ones); (38) boundaries of paleogeographic regions; (39) boundaries between facies zones in sedimentary basins; (40) main, and (41) secondary directions of sediment transport; (42) primary red color of rocks; (43) weathering crust.

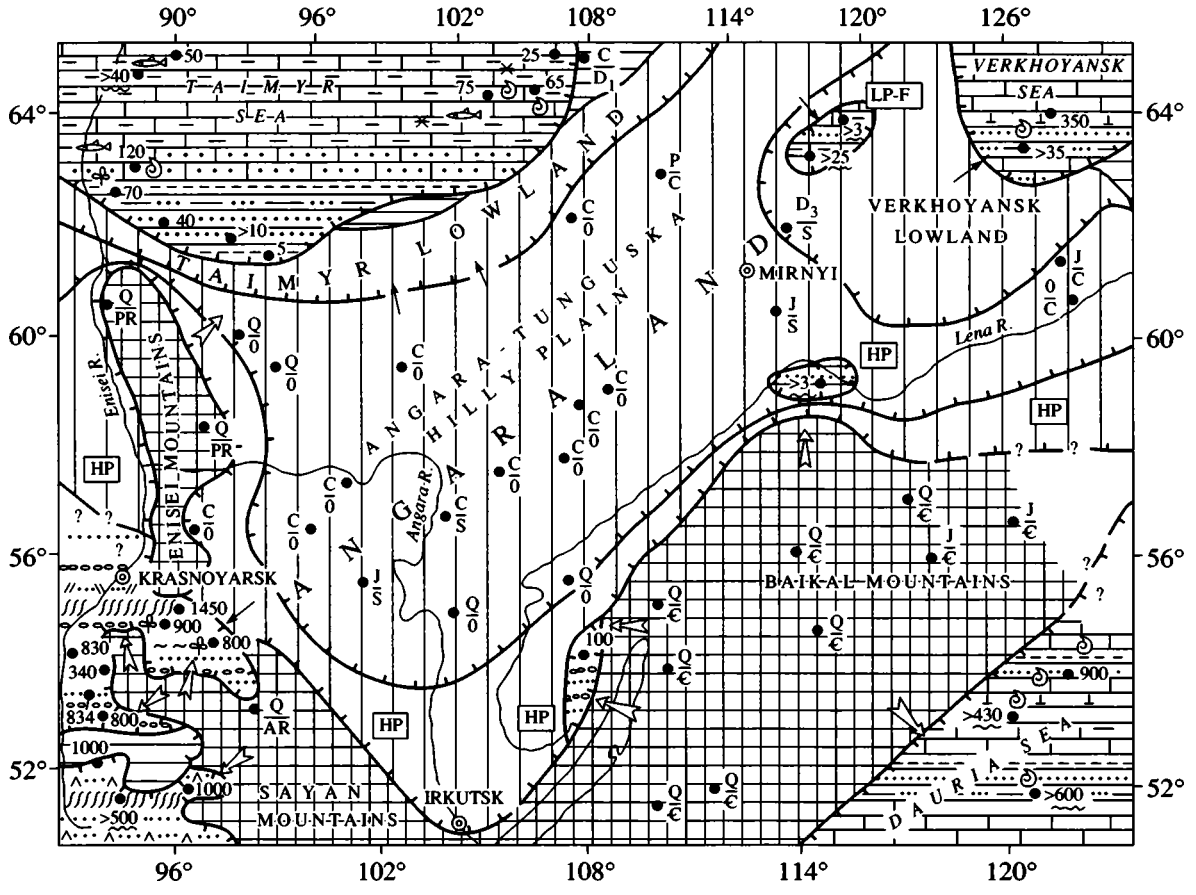

Fig. 2. Lithological–paleogeographic map of the southern Siberian platform, the latest Early–the earliest Middle Devonian (Emsian and Eifelian), symbols and scale as in Fig. 1.

plete Devonian section is that of the Rybnaya basin of the Middle Paleozoic sedimentation, where deposits of all three Devonian series are clearly distinguishable. Since the Poima–Biryusa, Pribaikal'skii, Ygyatta, and Kempendyai sedimentary basins originated later, the succession of Devonian deposits accumulated in these structures is incomplete.

At present, the Rybnaya basin is located east of Krasnoyarsk, being bounded in the south by fold structures of the East Sayan Ridge adjacent to the Siberian platform and in the north by the southern flank of the Enisei Range. The basin is as large as 110 × 330 km. This is one of the most complicated sedimentary basins, geologic history of which began in the Riphean at the formation time of the Sayan pericratonic trough. In the course of the Middle Paleozoic tectono-magmatic activation of the trough (epiplatform orogeny), the basin passed into the evolutionary stage typical of the marginal rift-related structure. Rocks of all the series of the Devonian System constitute the geological structure of the basin, the reference one for the southern part of the Siberian platform.

DEVONIAN SYSTEM (D)

LOWER SERIES (D_1)

Agul Formation (D_1^1)

Assaf'evka Formation (D_1^2)

LOWER–MIDDLE SERIES (UNDIVIDED)

Karymov Formation (D_{1-2})

MIDDLE SERIES

Pavlovka Formation (D_2)

UPPER SERIES

Kungus Formation (D_3)

The results of comprehensive study of reference sections in the Rybnaya paleobasin, most of which are compiled based on drilling results and description of boreholes, are summarized in the presented correlation scheme of Devonian deposits (Fig. 5). *The Agul Formation*, like its stratigraphic analogs in the East Sayan foothills (Byskar and Imir formations), unconformably rest on Lower Paleozoic and Proterozoic deposits, being confined to subsided blocks of the pre-Devonian basement. The formation is composed mostly of volcanogenic rocks, such as tuff, welded tuff, orthophyre,

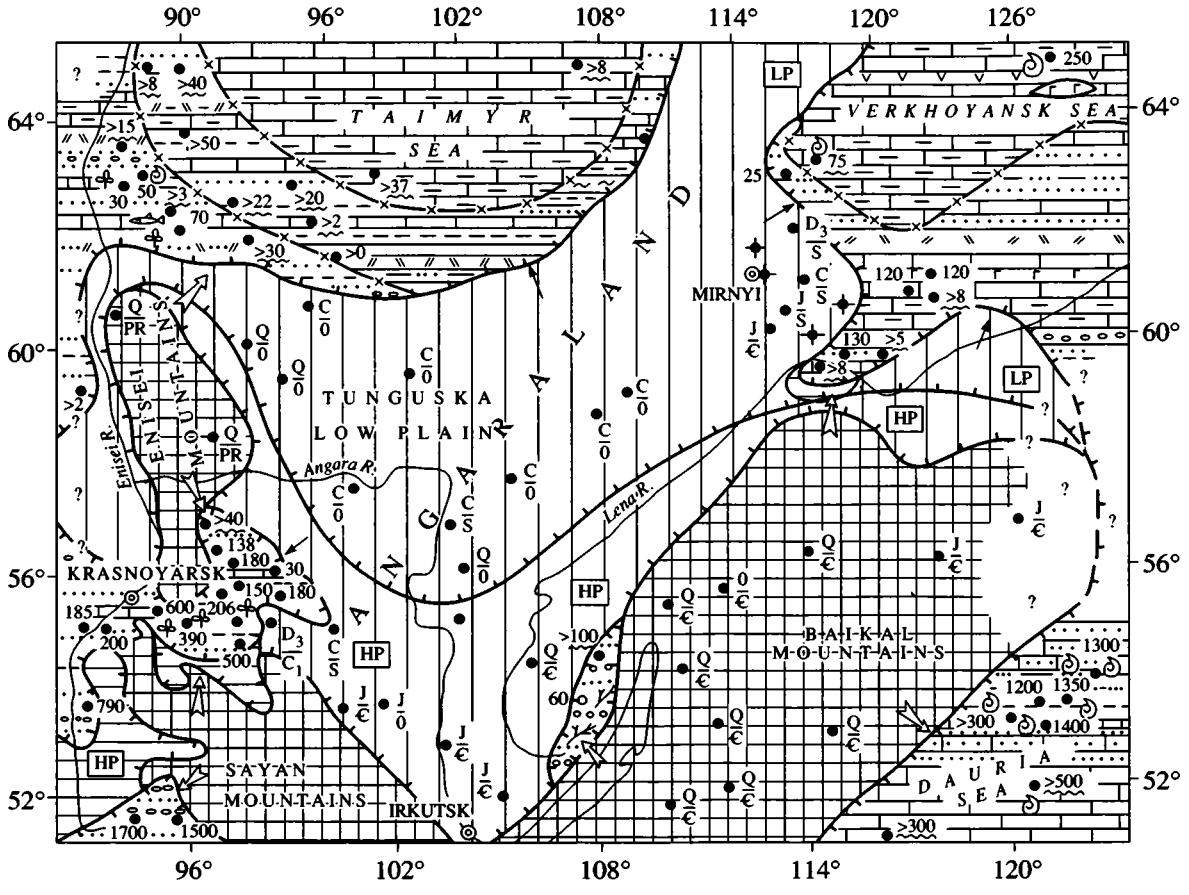

Fig. 3. Lithological–paleogeographic map of the southern Siberian platform, the Middle Devonian (Givetian), symbols and scale as in Fig. 1.

andesite-porphyrite, felsite, trachyte, trachyte-porphyrite, diabase, diabase-porphyrite, basaltic lavas and breccia. Typical volcanogenic-sedimentary and sedimentary deposits appear in the upper part of the section, being represented by tuff conglomerate and conglomerate, tuffstone and sandstone, and by tuff breccia. The Agul Formation is estimated to be 800 to 1000 m thick. The Early Devonian age of volcanogenic and volcanogenic-sedimentary deposits of the formation is established by radiological methods (Nozhkin and Smagin, 1979) and from stratigraphic relationships, as they underlie in places the deposits of fossil-bearing Assaf'evka Formation.

The Assaf'evka Formation was first distinguished by N.Ya. Kogan in 1948. The formation stratotype is located on the Rybnaya River right bank near the mill of the Assaf'evka village. I found additional exposures of the formation near the villages of P'yankovaya, Kubeinka (Yakhontov Log), and Kan-Okler. Almost everywhere, deposits of the formation rest with angular unconformity on the eroded surface of Cambrian and Precambrian rocks, and only in western areas they conformably overlie, though with a hiatus, the volcanogenic-sedimentary complex of the Agul Formation.

The basal brownish gray conglomerate of the formation is composed of pebbles that represent underlying Cambrian quartzite and quartz porphyry, and more rarely sandstone of the Koi Formation.

The overlying member consists of brownish gray, inequigranular sandstone with scattered pebbles in places. Sandstone often encloses intercalations of conglomerate similar to that of the basal bed.

Frequency of conglomerate interlayers decreases upward the formation section. Near the top, where they are missing, numerous interlayers of brownish gray siltstone appear instead. The formation is from 200 to 250 m thick.

Rare organic remains (psilophyte impressions) were found in the formation only near the village of Torgashino and in the vicinity of Sorokino Station. For instance, Bystritskiĭ (1964) found impressions of *Psilophyton goldschmidtii* Hall, *Psilophyton princeps* Dawson, *Drepanophycus spinaeformis* Goeppert, *Jenisseiphyton rudnevae* (Peresv.) Anan., *Protobarinophyton obrutschevii* Anan., and *Pectinophyton bipectinatum* Anan in grayish green medium-grained sandstone near the Sorokino Station. This plant assemblage is indicative of the Early Devonian age of host rocks. The

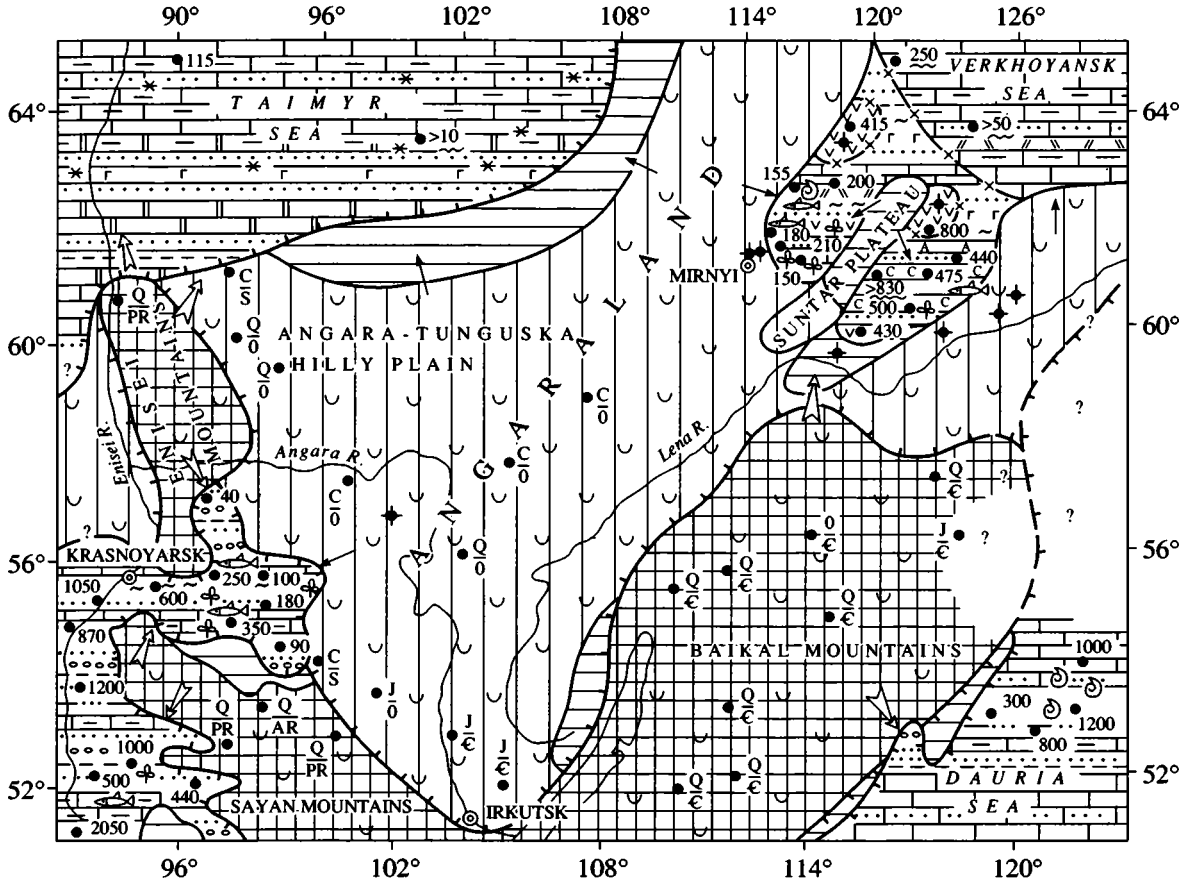

Fig. 4. Lithological–paleogeographic map of the southern Siberian platform, the Late Devonian (Frasnian and Famennian), symbols and scale as in Fig. 1.

boundary between the Assaf'evka and overlying Karymov formations is obscure, conventionally placed at the level of transition to frequent siltstone beds.

The Karymov Formation is represented by a sequence of siltstone, sandstone, and conglomerate red beds intercalated with volcanic lava sheets. By lithology, it is subdivided into lower and upper subformations.

The lower subformation is composed of siltstone, sandstone, gritstone, and conglomerate. Reddish brown siltstones and fine-grained sandstones occurring at the subformation base are cross-laminated, flaggy, showing ripple marks and desiccation cracks as a rule. The siltstone sequence is up to 500 m thick in total. The upper part of the subformation is mainly represented by sandstone with reddish brown siltstone, gritstone, and conglomerate interlayers. Also characteristic of the subformation are interlayers of greenish gray, fine-grained sandstone with plant remains. The total thickness of the subformation is 850 m.

It should be pointed out that boreholes, drilled near the village of Maganskoe, penetrated two volcanic lava flows separated by a sandstone member about 22 m thick and by cobble conglomerate with pebbles of granitic and volcanic rocks.

The upper subformation is composed mainly of psephitic rocks, such as gritstone and conglomerate. Sandstones and siltstones similar to those from the lower subformation are rare. The boundary between two subformations is unclear, placed arbitrarily at the appearance level of thick conglomerate members.

The total thickness of the Karymov Formation is 1450 m. V.M. Kovbasina and N.M. Petrosyan have found and identified *Jenisseiphyton rudnevae* (Peresv.) Anan., *Taeniocrada orientalis* Radcz., *Barrandeinopsis (?) antiqua* (Tschirk.) Radcz. in lower sandstones of the formation exposed in eastern outskirts of the village of Khaidak (a quarry near the road). I.P. Zhuiko and V.M. Kovbasina collected *Psilophyton burnotense* (Gilk.) Kr. et Weyl., *Uralia camdjalensis* Petr., *Dichophyton typicum* Radcz. on the left and right banks of the Sytik River in the neighborhood of the Sorokino Station. According to Kovbasina and Petrosyan (see in Andreeva *et al.*, 1962), the listed plant remains suggest the Early–Middle Devonian age of host deposits.

According to isotopic dating, basalts from the top of the Karymov Formation confirm the Middle Devonian

age of the unit (Nozhkin and Smagin, 1979). Hence, the formation age is assumed to be ranging from the Early (Emsian) to Middle Devonian (Eifelian).

The Pavlovka Formation described for the first time by M.A. Zharkov in 1956 was penetrated by boreholes drilled for structural study and mapping. Its deposits rest with a hiatus on the Karymov Formation and, in places, on older rocks. The formation is most completely exposed in the stratotype section along the Anzhe River near the village of Pavlovka. It is composed of alternating variegated sandstone (up to 75%), siltstone (15%), mudstone (5%), limestone and conglomerate beds (about 5%). In some places (near the villages of Zykovo and Verkhnyaya Esaulovka), sandstones are replaced by pebble conglomerates up to 100 m thick.

Comprising the reference horizons of limestone, hard gravelly sandstone, and sandstone with plant remains, the Pavlovka Formation is well traceable and distinct from other Devonian red beds. Its basal interval is composed of conglomerates intercalated with cross-laminated coarse-grained sandstone and gritstone lenses. Up the section, conglomerates grade into variegated medium- to coarse-grained polymictic sandstones with calcareous ferruginate cement. The sandstones show cross-lamination, ripple marks, and desiccation cracks. These rocks are overlain by a thick (70–80 m) terrigenous–carbonate member of alternating limestones, sandstones, and lumpy siltstones. Limestones are massive, fine-grained, dark to light gray, with characteristic inclusions of brown–red chalcedony. Individual limestone horizons range in thickness from 6 to 12 m. Sandstones are light gray to brown, inequigranular, polymictic, often calcareous, grading into arenaceous sandstones with chalcedony inclusions. Siltstones are lumpy, calcareous, and brown in coloration. The total thickness of the strata is about 160 m.

The higher horizons of the Pavlovka Formation are exposed in upper reaches of the Tugach River. These are reddish brown sandstones alternating with green, gray, and greenish gray carbonate varieties. Sandstones are intercalated with interlayers and lenses of lumpy calcareous siltstone and arenaceous limestones. Further eastwards, in the Kungus River basin, abundance of terrigenous rocks in the Pavlovka Formation decreases in contrast to increasing quantity of limestone beds. The formation is from 253 to 600 m thick.

Yanov (1955) reported on floral remains found in deposits of the Pavlovka Formation along the Rybnaya River near the village of Partizanskoe. E.F. Chirkova-Zalesskaya identified among them *Archaeopteris* sp., *Psilophyton* sp., and *Dicronophyton* sp. Similar fossils were encountered by V.T. Mordovskii and V.I. Ditmar in the upper part of the Ivashikha Formation (the stratigraphic analog of the Pavlovka Formation). In addition, A.N. Sanchuk found cirriped remains near the village of Shalinskoe, which were classed by N.I. Novozhilov with *Ctensenigma tuvaense* Nov. (Andreeva *et al.*, 1962).

Chirkova-Zalesskaya and Novozhilov attributed the Pavlovka Formation to the Middle Devonian (Givetian).

The Kungus Formation named by A.V. Lesgaft was distinguished by V.F. Dontsov in 1936. In the northern part of the Rybnaya structural–sedimentation zone, the Kungus Formation transgressively, with angular unconformity overlaps the Pavlovka Formation. In the southern part, it is in conformable relations with underlying deposits of the Pavlovka and Agul formations according to data on reference boreholes and mining trenches.

The formation is represented mainly by variegated (brownish to dark red, bluish green, light gray) calcareous–clayey rocks, such as marl, siltstone, and mudstone. These rocks are often intercalated with sandstone, gritstone, conglomerate, and lumpy limestone interlayers. Dark red calcirudite beds are typical for the formation. In outward appearance, they are similar to gritstones and conglomerates being composed of well-rounded clasts of light gray pelitomorphic limestone and brown-red marl, which are cemented by carbonate or clay–carbonate material. The cement/clasts ratio in calcirudites is variable, and over a short distance along the strike, they may smoothly or abruptly grade into calcareous–clayey siltstones or limestones.

Because of specific lithology of calcirudite marker beds, the Kungus Formation is reliably distinguished and traced almost through the whole territory of the Rybnaya basin. The formation is thickest in the southwestern part of the basin (400 m), being not more than 310 m thick in the northeastern part and about 200 m thick in the northwestern part, near Krasnoyarsk.

The formation age is determined based on numerous remains of fossil Placodermi. For instance, Khomentovskiĭ (1947) found *Osteolepis* sp., *Megalichtus* sp., *Bothriolepis* sp., *Dipterus* sp., and *Holophychius* sp. in siltstones and calcirudites near the villages of Novo-Mariinka, the Yagans River mouth (Agul River basin), Blagoveshchenskoe (Kungus River) and Charginskoe (Kan River).

In the southwestern of the Rybnaya basin, Mordovskii and Ditmar (1954) found fish remains *Bothriolepis* sp. and *Dipterus* sp. (identified by D.V. Obruchev) and phyllopods *Sphaerestheria celsus* Nov., *Brachystheria piskovishensis* Nov., *Asmussia vulgaris* (Lutk.) Nov., and *Trigonestheria timanica* (Lutk.) Nov. (identified by N.I. Novozhilov). According to Obruchev and Novozhilov, the identified fossils are typical of the Late Devonian.

HISTORY OF SEDIMENTATION AND PALEOGEOGRAPHIC CHANGES IN THE REGION

During the Early Devonian (Lochkovian Age), substantial changes in the structure, landscape, and cli-

matic environments that took place in the Angaraland predetermined further development of geological processes. The Silurian regression of the East Siberian epicontinental sea had occurred by that time and left behind a vast continental land (Fig. 1). The Angaraland was bounded by the Zaisan–Ob sea in the west (according to the present-day reference coordinates), by the Taimyr sea in the north, by the Verkhoyansk sea in the east, and by the Dauria sea in the south (Fig. 3). The triggered intense erosion and denudation in the Angaraland was accompanied by immense outburst of alkali-basaltic volcanism in the Sayan–Baikal mountain region.

The first Rybnaya basin of continental sedimentation that appeared in the Angaraland accumulated a thick (up to 1000 m) sequence of volcanogenic deposits unconformably resting on Lower Paleozoic and Proterozoic rocks. The deposits are confined to the subsided blocks of the pre-Devonian basement. After that, the 250-m-thick molassic sequence of cross-bedded sandstones and conglomerates, characteristic of which is coarse cyclicity, was deposited. Conglomerate beds of molasses grade into inequigranular polymictic sandstone and siltstone away from provenances. The composition of pebbly material depended on rocks of adjacent provenance. For instance, deposits of the Assaf'evka locality bear pebbles of Proterozoic rocks and volcanic complexes derived from the Khaidak Massif. Conglomerates and sandstones are poorly graded, belonging to typical facies of piedmont detrital funs of temporary streams and piedmont rivers with migrating stream channels. Streams transporting terrigenous material were powerful. In a single section, one can observe well rounded pebbles of rocks resistant to abrasion in association with angular fragments of rocks unstable against the abrasive factors. By the end of the Early Devonian, deposition of psephitic, poorly sorted material gradually ceased and gave place to accumulation of alluvial–lacustrine and shallow-water lacustrine facies. This facies complex up to 700 m thick is composed of graded, wavy cross-laminated sandstone, siltstone, and mudstone. Desiccation cracks up to 1.5 cm wide, convolution balls locally formed at the surface of quickly dried clay beds, and small current ripples typical of dying shallow lakes are observable on bedding planes of the complex.

The sections described above indicate that an intense downwarping of the paleobasin compensated by an intense influx of clastic material from provenances was a basic factor that determined the facies peculiarities of sediments deposited at the initial stage of the basin evolution. Red color of the molassic sequence suggests the oxidizing conditions in the paleobasin of sedimentation. Redeposited products of weathering present in the lower part of the Assaf'evka Formation (kaolinite in the sandstone and conglomerate cement) and absent higher in the section are indicative of the pre-Devonian laterite crust that existed in the provenance.

Lenticular sandstone interlayers with abundant mudstone pellets in the upper part of the formation point to rewashing of sediments at the time of freshwater basin shoaling. Phases of partial basin exsiccation periodically enhanced the mineralization extent of water in lakes, as it is evident from thin (1–2 mm) gypsum veinlets in siltstone and mudstone. Chemogenic calcite constituting 50–60% of cementing material in sandstones implies the alkaline environments of sedimentation in the basin during the deposition period of these rocks. The hot climate of the Early Devonian that replaced the Silurian humid climate became much drier and then turned into the arid one by the end of the Devonian. First terrestrial plants that appeared were psilophytes (*Psilophyton goldschmidtii* Hall., *Psilophyton princeps* Dawson, and others).

In the terminal Early–initial Middle Devonian period, the area of continental sedimentation on the platform part of the Angaraland became larger and sedimentation started in the Pribaikal'skii basin that also emerged at the platform margin near the ancient Baikal mountains (in the Baikal foredeep). The basin extended as a narrow band about 30 km wide from north to south over the distance of 400 km (Fig. 2). Conglomerate red beds with single boulders and interlayers of coarse-grained sandstone form the basal molasse sequence, the thickness of which increases up to 200 m in the NW–SE direction. Boulders and pebbles are composed of the Akitkan Group rocks, such as granitoids, felsite, quartz-porphyry, diabase, gabbro-diabase, vein quartz, limestone, and sandstone. The proportion of psephitic, poorly graded red beds markedly decreases upward the section, where coarse- to medium-grained sandstone varieties appear. Well-preserved imprints of fossil plants have not been encountered in the formation, although the rocks are enriched in coalified plant detritus.

In the Rybnaya basin, sedimentation was in progress within the area of deposition of the Assaf'evka Formation. During the period under consideration, a thick (900 m) sequence of conglomerate and sandstone red beds with gritstone lenses and single pebbles of volcanic and metamorphic rocks was deposited. The sequence characterizes continuation of the molasse deposition, but sedimentation in the eastern flank remote from piedmonts resulted in accumulation of fine-grained material (sand, silt, clay, and subordinate carbonates, which facilitated later on the formation of marl). Siltstone beds up to 500 m thick in total represent the main part of the sequence. In psephitic sedimentary cycles of the Lower–Middle Devonian, cross-lamination is usually unidirectional, and thickness of cross-stratified series decreases upward the section, where they are replaced by thin-laminated fine-grained sediments. These textural and structural features are typical of sediments of fluvial and flood-plain facies. The composition of clastic material (arkose and graywacke sandstones) points to erosion of volcanic and intrusive rock complexes. Conglomerates and sandstones with

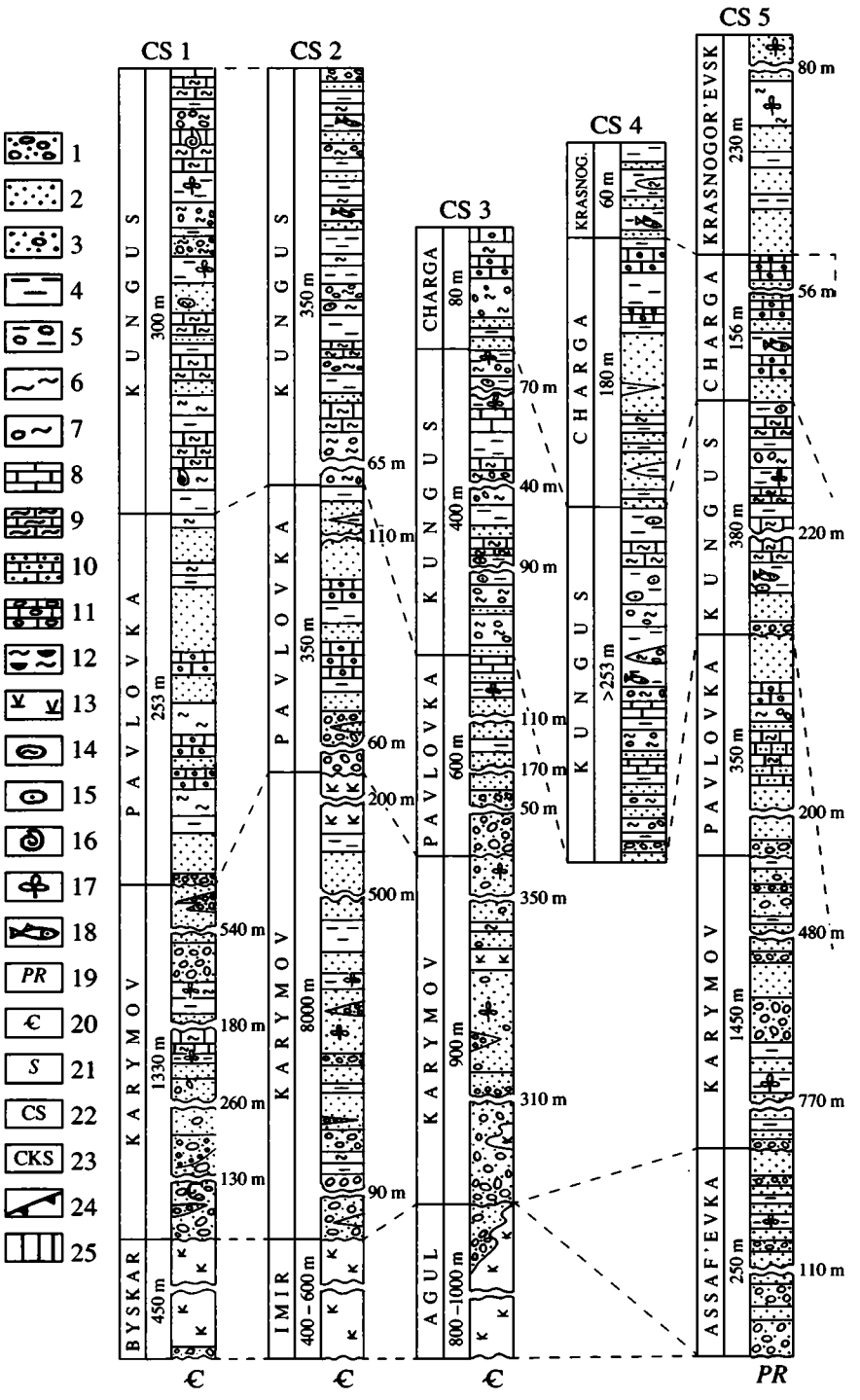

Fig. 5. Correlation scheme for reference core sections (CS) of Devonian deposits in the Rybnaya sedimentary basin: (1) conglomerate; (2) sandstone; (3) sandstone with dispersed pebbles; (4) siltstone; (5) siltstone with mudstone pebbles; (6) mudstone; (7) calcirudite; (8) limestone; (9) marl; (10) arenaceous limestone; (11) silicified limestone with chalcedony of a sealing-wax and red colors; (12) bentonite mudstone; (13) volcanic lava sheets; (14) mudstone pebble; (15) carbonate concretions; (16) fossil fauna, (17) flora, (18) fish fauna; (19) Proterozoic, (20) Cambrian, and (21) Silurian rocks; (22) key section; (23) composite key section; (24) boundary of sedimentary basin; (25) area of Devonian deposits.

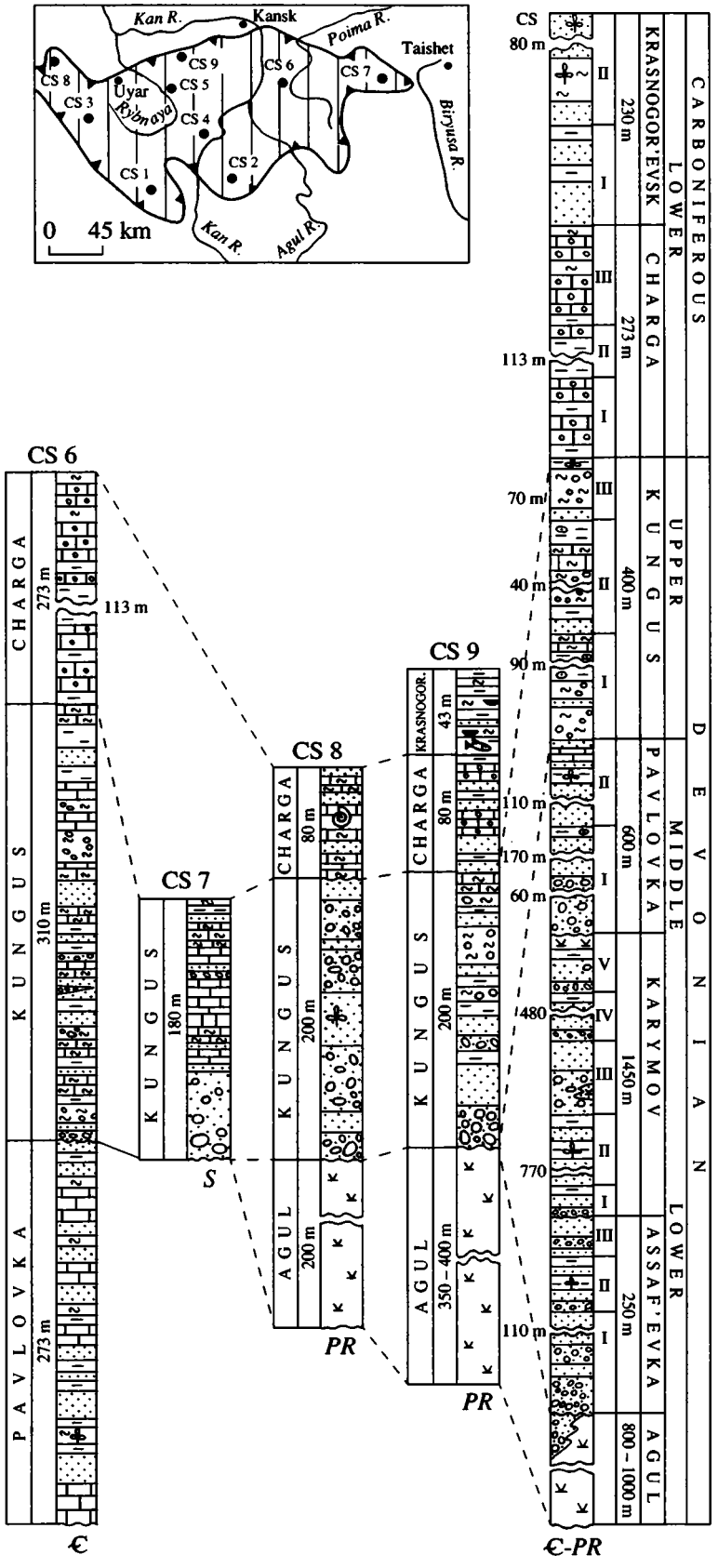
Kan R.
Kansk
Poima R.
Taishet
Birysa R.
Uyar
Rybnaya
CS 1
CS 2
CS 3
CS 4
CS 5
CS 6
CS 7
CS 8
CS 9
0 45 km
Kan R.
Agul R.
CS 6
CS 7
CS 8
CS 9
CS
CARBONIFEROUS
DEVONIAN
LOWER
UPPER
MIDDLE
LOWER
KRASNOGOR'EVSK
CHARGA
KUNGUS
PAVLOVKA
KARYMOV
ASSAF'EVKA
AGUL
230 m
273 m
400 m
600 m
1450 m
250 m
800 – 1000 m
310 m
180 m
200 m
80 m
43 m
350 – 400 m
80 m
113 m
70 m
40 m
90 m
110 m
170 m
60 m
480
770
110 m
PR
PR
Є
Є-PR

Fig. 5. (Contd.)

single pebbles and lenses of cross-laminated, inequigranular, polymictic, poorly graded sandstones represent sediments of lakes, deltas, and river streams. Fine-grained sandstones and siltstones with wavy and discontinuous horizontal lamination, which occur in association with deltaic facies, accumulated in an epicontinental lagoon-like basin. Silty mudstones and marls, as well as highly arenaceous, relatively thin limestones alternating with of deltaic sediments, seem to belong to facies of shallow small basins and could partially be deposited in subaerial settings.

Thus, the Sayan Mountains directly adjacent to the Rybnaya basin experienced intense destruction and erosion during in the Assaf'evka time, but the provenance of clastic material migrated in the southeastern direction to their spurs in the Karymov time. The area abraded during the Karymov time was of a medium height, as it is evident from sandy and silty size ranges of prevailing sediments despite the provenance proximity. The hot climate around the water basin stimulated the development of plant community that included *Jenisseiphyton rudnevae* (Peresv.) Anan., *Taeniocrada orientalis* Radcz., *Barrandeinopsis* (?) *antiqua* (Tschirk.) Rad., *Psilophyton burnotense* (Gilk.) Kr. et Weyl., *Uralia camdjalensis* Petr., and *Dichophyton typicum* Radcz.

The Angara–Tunguska hilly plain occupied a vast area of the Angaraland, being connected with high piedmont plains, on the one side, and turning into low plains near the Taimyr and Verkhoyansk seas. Denudation and erosion of Silurian, Ordovician, and Cambrian deposits were in progress everywhere.

In the Middle Devonian (Givetian), the area of sedimentation substantially increased in the Rybnaya basin, where chemogenic limestone, marl, and other rocks accumulated. Configuration of the basin considerably changed as compared to the previous one. If the basin formerly represented a narrow linear trough, then it became of nearly isometric shape, 230 km long and up to 110 km wide, in the Pavlovka time, when the growth of the East Sayan Mountains caused the uplift of adjacent basin areas and ingression in the northeastward direction, which resulted in the basin widening (Fig. 3). The resultant sedimentation is recorded within a considerable southwestern area of the Siberian platform. In central regions of sedimentation, which were remote from the main provenance, the accumulation of lagoonal, deltaic, alluvial, and lacustrine facies was in progress.

The sedimentation area in the Pribaikal'skii basin also increased, though in the northwestern direction only. This basin was about 400 km long and up 100 km wide at maximum. It accumulated sandy material with dispersed quartz pebbles. Terrigenous sandstone grains are subangular, poorly sorted, set in the contact–porous, ferruginate–calcareous cement. The short-term tectonic activation in Baikalides was responsible for deposition of the 25-m-thick sequence of intraformational small-pebbled conglomerates. Sedimentary fill of the basin characterizes geological activity of temporary streams, rivers, and lacustrine sedimentation. Sometimes, rivers running from the ancient Baikal mountains and supplying the bulk of clastic material were of a great transporting energy, able to carry boulders.

In the Middle Devonian, three other sedimentation basins: the Poima–Biryusa, Kempendyai, and Ygyatta, originated in the Angaraland platform, the main area of which was still occupied by the Angara–Tunguska hilly plain. The formation of the Poima–Biryusa basin was determined by deep-seated faulting and lateral displacements in the basement of the southern Siberian platform. The basin represented a linear N–S trending structure 80 km wide and 480 km long.

Owing to transgression from the Rybnaya bay, waters penetrated into the Poima–Biryusa sedimentary basin through the Enisei strait. The strait exsiccated sometimes, and the Poima–Biryusa basin turned into an ordinary lagoon. The Okino–Vikhorev flexure was a principal tectonic structure that limited the basin size. Extending from north to south, it restricted the ingression in the eastward direction.

Genetically, the Okino–Vikhorev flexure is related to basement faulting that is reflected as well in sedimentary cover, the Devonian deposits included. A chain of ridges composed of Ordovician and Silurian rocks extends at present along the western limb of the flexure. The ridges characterize the complicated differential block displacements of the platform basement in the area in question.

The Pavlovka Formation is up 180 m thick in the Poima–Biryusa basin. Its basal conglomerates unconformably rest on Lower Paleozoic rocks. Higher in the section, they grade into light gray sandstones with interlayers of limestone, siltstone, calcareous mudstone, and marl. Nodules of dark red cherts and gray chalcedony are characteristic of limestones in the middle part of the formation. The Middle Devonian age of these deposits is established based on isotopic analysis of chemogenic limestone (Akulov *et al.*, 1990).

The Ygyatta and Kempendyai sedimentary basins jointly represented a bay of the Verkhoyansk sea. The bay area was 200 × 320 km, and Middle Devonian sediments deposited here are about 130 m. According to Masaitis *et al.* (1975), the Ygyatta and Kempendyai basins form a part of the Patom–Vilyui aulacogen and belong to rift-related structures. They are separated by the Suntar uplift bounded by deep-seated faults with displacement amplitudes up to 2–4 km (Grishin *et al.*, 1987). Gray and red clay limestone, red–brown dolomite, crimson–red marl, siltstone and sandstone (a variegated sequence) are the most representative rocks deposited at that time in the bay. Basal conglomerates and dolomites (in places) unconformably rest on Lower Silurian deposits with a stratigraphic hiatus. Thin flood basalts and interlayers of synsedimentary tuffs are indicative of volcanic activity near the bay.

I.P. Tabachnikova who studied collection of M.V. Mikhailov found a spore assemblage similar to Middle Devonian (Eifelian–Givetian) assemblages of Bashkiria and western Urals. In addition, M.V. Mikhailov, V.N. Tikhii, V. Vl. Menner, G.S. Fradkin, and K.E. Kolodeznikov collected in various periods the remains of corals *Ceratophyllum* aff. *typus* Gurich., brachiopods *Emanuella takwanensis* (Kays.), pelecypods *Eoschizodus* sp., *Paracyclas* sp., trilobites *Dechenella* cf. *verneuili* (Barr.), foraminifers, gastropods, and crinoids of the Middle Devonian age (Kolodeznikov, 1982).

In the Late Devonian, the Rybnaya basin was ellipsoidal in shape. Its longer axis (up to 300 km) extended from east to west. The basin merged sometimes with the Poima–Biryusa lagoon into a single basin that accumulated red marl, siltstone, and mudstone beds. Grayish green limestone and calcirudite composed of well-rounded pebbles of red marl were also deposited here.

The Rybnaya–Biryusa basin of an irregular shape occupied an area of more than 120000 km^2. Regular shoaling in the basin led to appearance of isolated lagoons and, in places, of numerous lakes and swamps. The relief in the provenance (area of the East Sayan and Enisei mountains) was slightly dissected, as it follows from absence of psephitic deposits in the 400-m-thick sequence of terrigenous–carbonate deposits. Gray well-sorted quartz sandstones are the widespread deposits of that time. Imprints of fish remains *Bothriolepis* and *Dipterus* were found in the southwestern part of the Rybnaya basin along with phyllopods *Sphaerestheria celsus* Nov., *Brachystheria piskovishensis* Nov., *Asmussia vulgaris* (Lut.) Nov., and *Trigonestheria timanica* (Lut.) Nov. (Mordovskiĭ and Ditmar, 1954).

Intense tectonic displacements in the platform basement along deep-seated fault zones resulted in activation of the Patom–Vilyui aulacogen and in a further growth of the Suntar Uplift that divided the unified Ygyatta–Kempendyai bay into two parts (Fig. 4). The tectonic activity was accompanied by an intense volcanism, which resulted in deposition of the 415-m-thick sequence of flood basalts in the Ygyatta basin and in appearance of thin volcanic flows (up to 15 m) in the Kempendyai basin. The Ygyatta and Kempendyai basins sometimes lost connections with the Verkhoyansk sea (because of the eastward regression) and turned into two vast lagoons nearly equal in size, each covering the area of 60000 km^2.

It is significant that sedimentation in two lagoons was different, as it is evident from the thickness and composition of their sediments. Being up to 200 m thick in the Ygyatta basin, the accumulated sediments are 830 m thick in the Kempendyai basin. Rock salt, dolomite, tuff, anhydrite, and marl predominate in the latter, whereas variegated dolomite, limestone, mudstone, marl, ash tuff, and gypsum interlayers prevail in the Ygyatta basin.

On the left bank of the Vilyui River, 8 km downstream of the Verkhnii Menk Settlement (the Ygyatta basin), N.A. Flerova found remains of the Late Devonian Placodermi *Bothriolepis* sp. A diverse spore assemblage was established in red–brown mudstone of the formation lower part. The assemblage includes typical Famennian taxa *Acanthotriletes tenuispinosus* Naum. Var. *famenesis* Naum., *Lophotriletes atratus* Naum., and others associated with species *Archaezonotriletes opirarus* Naum., *Hymenozonotriletes radiatus* Naum., and *Lophozonotriletes tylophorus* Naum. Characteristic of Frasnian deposits in the Russian platform (Kolodeznikov, 1982).

In the Angaraland, the main part of which was occupied by plains of different types, the intrusion of diamond-bearing kimberlite took place. In areas of kimberlite pipe calderas, the rocks experienced erosion. It was the time, when the Mir kimberlite pipe intruded the hilly plain flooded sometimes by sea in the Ygyatta bay area (Brakhfogel and Koval'skii, 1974).

SUCCESSION OF PALEOCLIMATIC CHANGES IN THE ANGARALAND

According to reconstructed trajectory of the apparent pole wandering based on paleomagnetic data, the Siberian continent crossed the equator and entered the Northern Hemisphere in the Silurian (Khramov *et al.*, 1982; Pecherskii and Didenko, 1995; Smethurst *et al.*, 1998; Pavlov and Galle, 1999).

Based on the principle of actualism, it is possible to assume that the Devonian climatic zones were similar to present-day ones. Coordinating them with the trajectory of moving Siberian continent, we can trace the successive drift of continent across paleoclimatic zones. The results show that Siberian continent was in southern equatorial and subequatorial climatic zones nearly throughout the entire Silurian Period. In the Devonian, the continent entered the tropical zone of the Northern Hemisphere and then drifted across the subtropical zone during the Early Carboniferous. Surprisingly, but exactly this succession of climatic changes is inferable based on lithological indications of the paleoclimate, which are recorded in the Angaraland. The tropical arid climate that existed on the Angaraland throughout the Devonian was responsible for deposition of red beds and associated gypsum, rock salt, carbonate, and anhydrite. Desiccation cracks and convolution structure characteristic of sedimentation under arid climatic conditions are traceable throughout the study region. In addition, Akulov and Kashik (1999) studied the nature of the Devonian rocks coloration in the Siberian platform. Their thermodynamic calculations and modeling of goethite dehydration and transformation into hematite showed that this process is feasible only under hot arid conditions typical of the tropical zone.

CONCLUSIONS

Summarizing the considered data, it is necessary to emphasize the following.

(1) During the Devonian, the continental part of the Siberian platform (the Angaraland) was occupied by plains of different types with five, the Rybnaya, Poima–Biryusa, Baikal, Ygyatta, and Kempendyai basins of continental sedimentation.

(2) Sedimentary complexes of different age accumulated in sea bays, lagoons, and lakes. Rivers and temporary mountain streams transported the bulk sedimentary material into the basins, but in sea bays transformed in lagoons, sedimentation was of combined terrigenous–chemogenic type.

(3) The East Sayan, Enisei, and Baikal mountain systems were main provenances of terrigenous material. The tectonic–magmatic activity was most intense in the East Sayan Mountains. Their growth caused migration of sedimentation areas in the Angaraland interiors.

(4) Middle Paleozoic climatic changes in the Angaraland were determined by a gradual replacement of the hot and humid Silurian climate by the hot arid climate of the Devonian time, and then by the warm arid environments of the Early Carboniferous epoch. The established succession of climatic changes is consistent with paleomagnetic data suggesting the Siberian continent drift across paleoclimatic zones of Southern and Northern hemispheres during the Paleozoic.

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Serpukhovian Stage Stratotype in Zaborje Quarry (Moscow Region). Part II. Subaerial Exposure Profiles and Cyclicity

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Abstract—Muratovka Paleokarst at the top of Venevian and Dashkovka Paleosol at the top of Steshevian are described in Zaborje Quarry. The Dashkovka profile is classified as a durisol with near-surface calcrete hardpan. The pre-Moscovian subaerial profile developed upon Protvian limestones is briefly described. Nodules in nodular Venevian limestones appear to be the product of early freshwater lithification, and their earlier intraclastic interpretation has to be rejected. In addition to previously recognized large-scale cyclicity, the Zaborje section exhibits smaller-scale deepening–shoaling cyclites. The lower Serpukhovian (Tarusian–Steshevian) cyclite consists of 4 or 5 smaller-scale cyclites. Preserved in Zaborje lowermost part of the upper Serpukhovian cyclite contains at least one smaller-scale cycle.

Key words: Serpukhovian Stage, diagenesis, subaerial exposure profiles, cycles.

INTRODUCTION

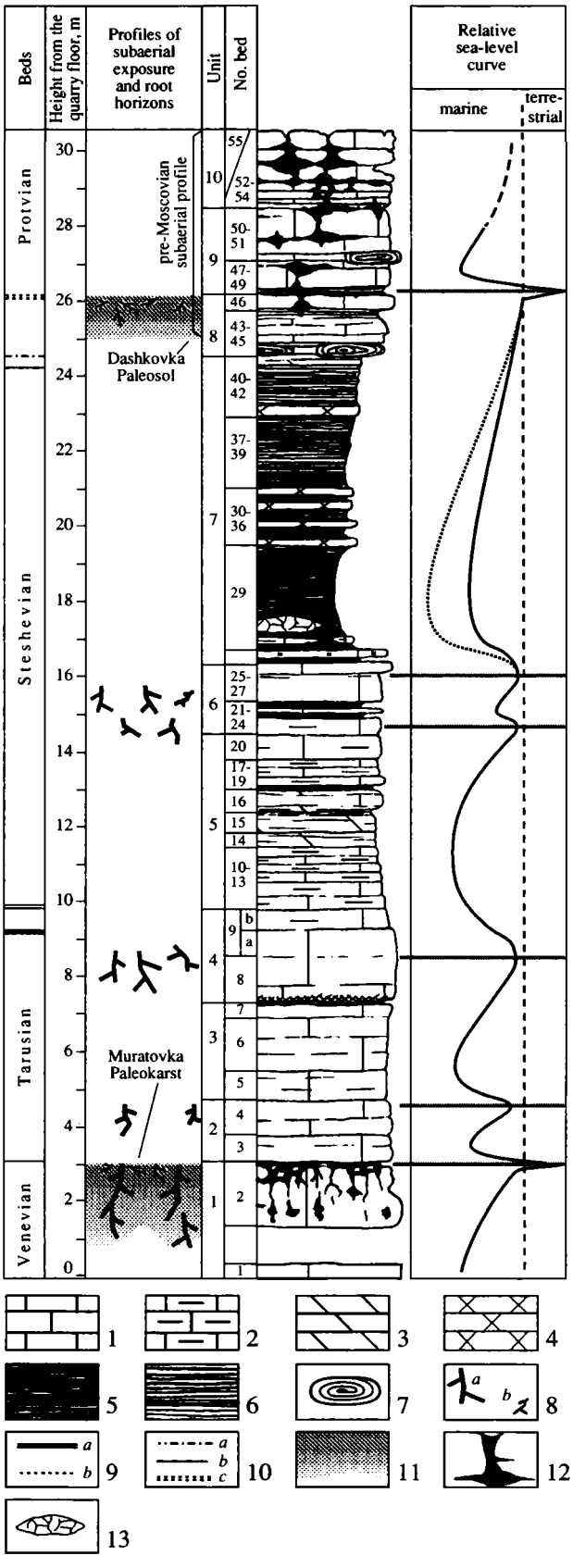
This paper continues the study of the section Zaborje (Kabanov, 2003). Facies characteristics, succession of shoaling levels, and subaerial exposure profiles allow recognition of the two-order cyclicity. Muratovka Paleokarst and Dashkovka Paleosol in Zaborje are important marker horizons that should be treated as boundaries of large subdivisions. In addition, the signatures of freshwater diagenesis occur in many limestone and marl beds in the lower carbonate part of the section. Shoaling levels without appreciable subaerial exposure surfaces are suggested by lycopsid stigmaria and other kinds of root traces (Kabanov, 2003). In the present paper, the mottling of the Okaian–lower Serpukhovian limestones is also attributed to the products of freshwater diagenesis.

RELATIVE SEA-LEVEL CURVE

The relative sea-level curve (Figure) corresponds to the "second-order cyclicity" of Shvetshov (1938) and, with modifications, to the seven- and nine-order rhythmicity of Makhlina *et al.* (1993). Two large-scale cyclites are recognized. The Tarusian–Steshevian interval corresponds to the "sixth Dinantian cycle" of Shvetsov (1938) and the Zaborje seventh-order rhythm of Makhlina *et al.* (1993). The Protvian Regional Substage belongs to the large-scale cyclite treated by Makhlina *et al.* (1993) as the upper Serpukhovian Staryi Beshev seventh-order rhythm. The two large cyclites are separated by the Dashkovka Paleosol.

Smaller-scale sea level fluctuations (Figure) correspond approximately to the ninth-order rhythms of Makhlina *et al.* (1993). Author's data do not support existence of the Tarusian and Steshevian eight-order cyclites (or rhythms) proposed by Makhlina and coauthors. The boundaries of four lower Serpukhovian small-scale cyclites are indicated by rooted horizons within shallow-marine beds. The lower Tarusian cyclite corresponds to Unit 2, and the upper Tarusian to units 3 and 4 excluding its upper Bed 9b. The lower Steshevian small-scale cyclite embraces the upper half of Unit 4, Unit 5, and part of Unit 6 (Figure). An additional thin cyclite may exist within Unit 6 if the fourth rooted horizon will be divided in two closely spaced rooted horizons. Three (or four) cyclites in the Tarusian–lower Steshevian part of the section are characterized by low facies contrasts. The rooted horizons are not accompanied by distinct subaerial exposure features (Kabanov, 2003). It seems that moderate or even slight sea level falls were insufficient to appreciable facies changes, but allowed colonization of the shallow seafloor by a type of vegetation ecologically similar to modern mangrooves.

The upper Steshevian cyclite (units 7 and 8) is capped by the Dashkovka Paleosol. This distinct cyclite begins with shallow-marine beds 27 and 28. The latter grade upward into soft black shales that must represent an offshore cyclite core. Further regression is expressed in shift from the soft shale to palygorskite hard shale. The core and regressive parts contain layers and lenses of early diagenetic dolomitic marl. The terminal regressive laminated intertidal rock is capped by the Dashkovka Paleosol (Kabanov, 2003). The shallow-water lagoonal interpretation of soft shales in the core of the upper Steshevian cyclite (Osipova and Belskaya, 1965a; Makhlina *et al.*, 1993) implies again low-amplitude sea level rise (solid line in the figure), which may



The Zaborje section. Subaerial exposure profiles, rooted horizons and the relative sea-level curve. (1) Limestone; (2) argillaceous limestone; (3) calcareous marl; (4) dolomitic marl; (5) soft shale; (6) palygorskite shale; (7) cherts; (8) root traces and stigmaria: (*a*) large, (*b*) small; (9) Tarusian/Steshevian boundary: (*a*) after Shvetsov, 1948 and Makhlina *et al.*, 1993, (*b*) after Kabanov, 2003; (10) Steshevian/Protvian boundary: (*a*) after Makhlina *et al.*, 1993, (*b*) after Barskov and Alekseev, 1979, (*c*) after Kabanov, 2003; (11) subaerial exposure profiles; (12) pre-Moscovian subaerial complex; (13) barite–limonite concretions.

be the lowest for the Serpukhovian time. The sharp facies contrast in this model was explained by an increased isolation of a gulf-like area in the southwestern part of the East European epicontinental sea and restricted to this area increased terrigenous influx (Osipova and Belskaya, 1965a). The first part of this work (Kabanov, 2003) offers another interpretation. The soft black shales may have accumulated in relatively deep-water settings below the depth of normal aeration. The density stratification of the water column may have taken place instead of isolation within the lagoon. In this case, deepening could be significant and even the maximal for the early Serpukhovian time (dotted line in the figure).

The lower part of the Protvian Regional Substage preserved in the section contains at least one distinct smaller-scale cyclite. Its most offshore part (beds 48–49) is strongly shifted to the base and separated from the Dashkovka Paleosol only by the thin palygorskite clay of Bed 27. Unfortunately, strong karstification masks details of overlying Protvian beds. The shallower-marine facies, however, are detected in thin sections up to Bed 54. If this part of the section belongs to the same cyclite, its regressive part is at least three times thicker than the transgressive plus core parts. Such behavior of the sea-level curve is different from that in the lower Serpukhovian. Apparently, the amplitude of sea-level fluctuations increased substantially in the earliest Protvian or in late Steshevian time, which correspond to evidences of aridization (Osipova and Belskaya, 1965a) and increased stronger movements of air masses as suggested by appearance of high-energy sedimentary textures (Kabanov, 2003). Possible relation of these local changes to global climatic differentiation in response to Gondwanan ice sheet development is yet to be investigated.

SUBAERIAL EXPOSURE PROFILES

Muratovka Paleokarst

This profile is named after the Muratovka quarry (Kaluga Region) where this horizon is well expressed (Osipova and Belskaya, 1965a). The "mottled rhizoid limestone" in the uppermost Venevian was first described in Aleksin by Shvetsov (1932) who called it "the stigmarian phase of shoaling" (Shvetsov, 1938). Osipova and Belskaya (1965b) have traced this horizon

across the southern flank of the Moscow syneclise at distance over 300 km and stressed its significance as a regional marker of the Venevian top. In 2000, the author detected a paleokarst profile matching the "rhizoid limestone" at the base of the Zaborje section.

The profile developed upon Bed 2 consists of the hard limestone 2a, the powdery horizon 2b, and the dark clay crust 2c (Plate I, fig. 1). The main part of the profile (2a) is a hard pale gray indistinctly grainy limestone with dark irregular mottles of cementation-recrystallization. Coalesced mottles just under the top of 2a give at places brownish gray microcrystalline appearance of the limestone. The rock is brecciated and pierced by dissolution vugs and channels (Plate I, fig. 1). Although enlarged by dissolution, the channels retain their primary structure. They can branch, meander, or form bunches connected with the tongues of the overlying powdery horizon. The channels must represent root voids ("rhizoid traces" in previous works) and mark the *first rooted horizon*. Many root voids penetrate down to over 1.2 m from the top. The limestone is irregularly bleached and loosened around individual dissolution cavities and their concentrations. Degree of dissolution and loosening is generally increasing to the limestone top. The signs of subaerial diagenesis in thin sections also become more pronounced upward. These are the laminae of vadose micritization (Plate I, fig. 2), crystalline veins, bladed and equant sparite cement, recrystallization, and dissolution cavities (Plate I, figs. 3–5). The dissolution voids are sometimes outlined by micrite linings and filled by the geopetal sediment that is recognized as vadose silt (Esteban and Klappa, 1983).

The hard limestone 2a reaches at places the top of the subaerial profile. Between these elevations, the tongues of loose powdery material of 2b extend to the depth of 0.3–0.4 m. The hard limestone is overlain by the powdery horizon (2b), which is mostly 5–15 cm thick and composed of hard limestone nodules and bleached powder calcite. The nodules are potato-shaped, 3–4 cm across, darker than the powdery matrix, and pierced by curved tubules with corroded or relatively smooth walls. Freshly broken nodules expel weak bitumen or sulfur smell. The primary marine limestone textures in nodules are strongly recrystallized and cut by different vadose textures including the tubules (Plate I, figs. 4, 5). Sometimes a main tubule up to 5 mm in diameter penetrates the long axis of the nodule, and subordinate tubules around 1 mm in diameter run at different angles to the main one. The orientation of main tubules is generally horizontal. Concentrations of tubules also occur in the walls of lithified blocks at the contact with powdery material (Plate I, fig. 3). These near-surface tubules may have been left by the roots of small plants. The enhanced cementation/crystallization responsible for nodule formation is related to root activity, as suggested by presence of root tubules inside nodules. The formation of cement sheaths around roots is a widespread phenomenon in $CaCO_3$-enriched substrates and hot climate settings (Klappa, 1980). The uptake of water by roots from the Ca^{2+}-rich ground solution sharply increases concentration of this ion, which may result in precipitation of calcite cement.

The top of Bed 2 is marked by a discontinuous crust of dark brownish gray to almost black clay 1–3 cm in thickness (2c in Plate I, fig. 1). Some limestone nodules from 2b pierce the dark clay crust 2c and project at the top of subaerial profile. These projections are draped by thin dark film.

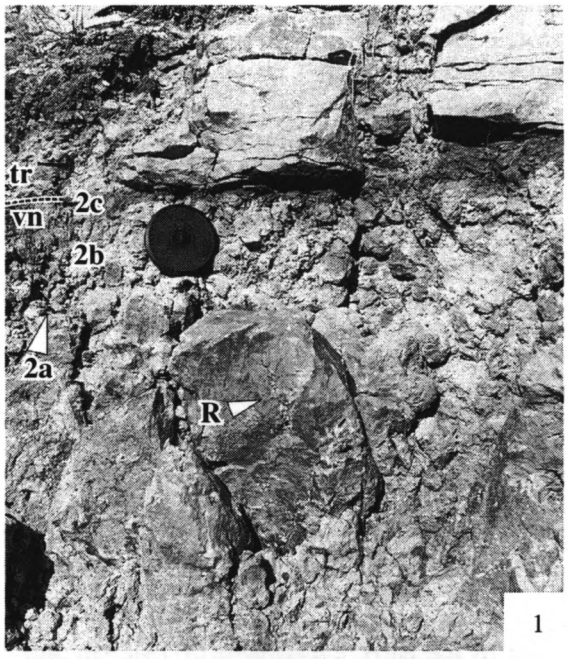
Abundance of small karstic cavities, which largely inherit root paths, allows the Muratovka profile to be referred to the type of the small-scale karst that is known as phytokarst (Folk *et al.*, 1973; Esteban and Klappa, 1983), microkarst, or pseudomicrokarst (Freytet and Plaziat, 1982). Soil features in the Muratovka profile are poorly developed. The powdery horizon 2b and the dark clay 2c may represent the original loose material accumulated between the blocks of cemented and karstified limestone. The Muratovka profile, which is 1.5–2.0 m thick and does not show significant dissolution relief, may correspond to the initial stage to the early karst stage of Esteban and Klappa (1983).

The powdery material of 2b resembles the "moon milk" (Esteban and Klappa, 1983), which is a loose product of parent rock degradation accumulated in depressions and cavities of the near-surface karst. On the other hand, its formation may be related to younger interstratal karstification. The latter is suggested by that the basal 1–2 cm of overlying Bed 3 is unevenly weathered up to the loose state resembling the underlying powdery material. At places, Bed 3 is broken by vertical brecciated zones with loose weathered material (Plate I, fig. 1), and its base shows dissolution relief. Therefore, it cannot be ruled out that loose and powder material is entirely the product of interstratal karstification. Apparently, this late alteration has affected both the top of Venevian limestones and the transgressive base of Tarusian limestones.

Dashkovka Paleosol

The profile is named after the Dashkovka quarry of palygorskite clay in the Serpukhov District where this paleosol is exposed. Pedogenic features in this profile are well developed, while vadose dissolution features are practically indistinguishable against the background of the pre-Moscovian karstic terrane. The 0.3–0.5 m-thick Bed 46 seems to be almost completely pedogenic. In the field it is seen as yellow to pale brown aphanic limestone with indistinct small-scale brecciation, massive in the lower half and grading upward into platy marly rock (Plate II, fig. 2). Brecciated appearance is determined by twisting short fissures and fenestrae up to 3–5 cm long and intertangled tubules 1–2 cm across. The majority of fenestrae and tubules are filled with light soft carbonate. Hollow fenestrae are uncom-

Plate I



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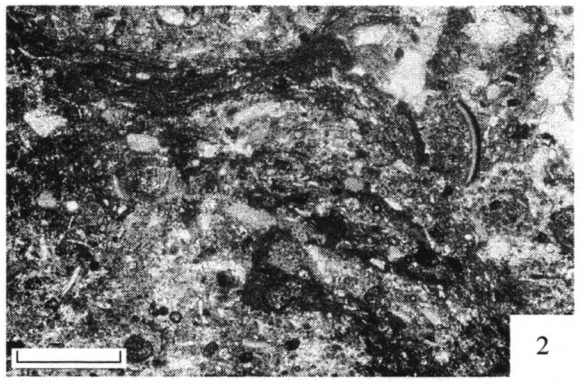

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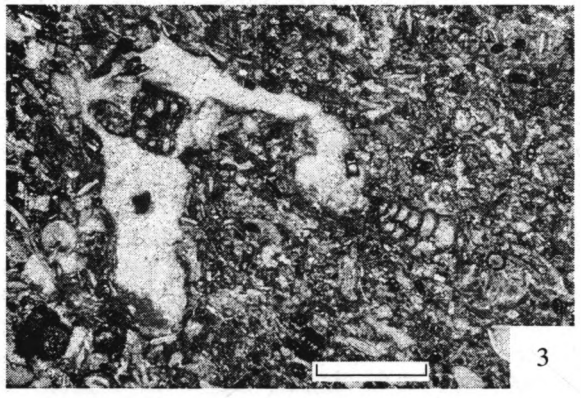

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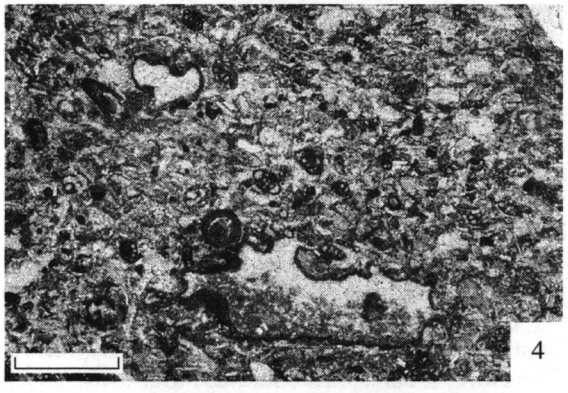

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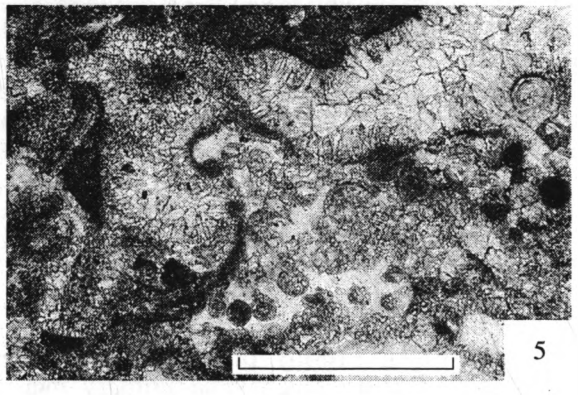

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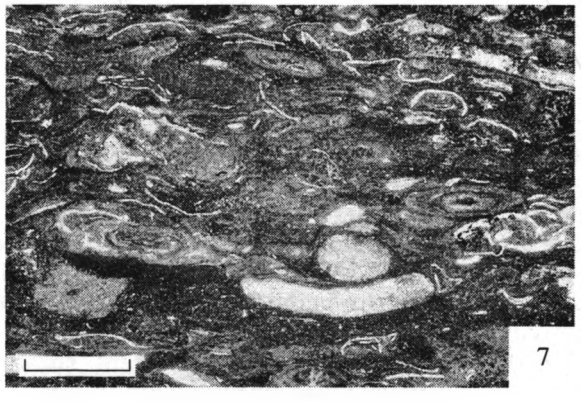

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mon. Tubules and fenestrae are sometimes distinctly superimposed upon earlier generations of similar structures that are highly degraded. Sometimes the tubules are branching (Plate II, fig. 2). Their root origin leaves little doubt. These root traces constitute the *fifth rooted horizon* of the Zaborje section.

The remnants of primary textures (the lenses of ostracod–molluscan rudstone) are present only at the base of Bed 46. A thin section from one of these lenses shows the highest diversity of pedogenic textures (Plate II, fig. 3–5). The soil carbonate cementing and replacing the primary rudstone texture consists of two phases (Plate II, figs. 4, 5). The first phase is represented by cryptocrystalline (1–3 μm) homogeneous to vaguely laminated brown micrite (Plate II, fig. 3). This phase represents calcite cutans (calcans) in the form of irregular overgrowths and thick micrite bridges that are locally coalesced into micritic mottles. The second phase is represented by cloudy lighter microsparite replacing the first phase with sharp boundaries. The microsparite replacements tend to develop inside the calcans of brown micrite (Plate II, fig. 4). The microsparite is looser and consists of idiotopic rhombohedra 10–20 μm in diameter and degraded micrite spots (Plate II, fig. 7). Other pedogenic structures are lacy structures (Plate II, fig. 3), aggregates of small (approximately 100 μm) peloids suggesting presence of soil infauna, pedotubules of at least two kinds, and large complex crudely concentric tubular structures. Pedotubules of the first kind are small channels approximately 0.5 mm in diameter that lack distinct signatures of root or animal burrow origin. Such tubules comprise the bulk of the pore space in the thin section (Plate II, fig. 4). Tubules of the second kind are 70–100 μm in diameter. They penetrate the periphery of complex tubular structures (Plate II, figs. 4, 5). These latter represent a kind of rhizoliths (Klappa, 1980). One oblique cross section of a complex tubular structure over 1.5 cm across shows the core of structureless loose microsparite and the thick peripheral layer consisting of compact microsparite pierced by tubules of the second kind. These latter might have been left by rootlets.

The upper part of the Bed 46 lacks primary sedimentary textures. The rock is dominated by brown uniform micrite. Microsparite occurs in lesser quantity. It fills up the fenestrae and root tubules. Peloids and pedotubules of the first kind are also present (Plate II, figs. 6, 7). Such texture corresponds to dismicrite of Folk (1959). A biogenic structure found in the thin section from the bed top (Plate II, fig. 6) might be a calcified plant tissue. Thin sections from the underlying Bed 45 lack distinct pedogenic structures. The subaerial diagenesis related to Dashkovka exposure may be represented by dense equant sparite cement (Plate I, fig. 6), pre-compactional lithification, and recrystallization of the rock.

The argillaceous limestone of Bed 46 is uncompacted except for stylolites, despite the non-carbonate admixture should trigger compaction even under minor lithostatic pressure (Tucker and Wright, 1990). This means lithification prior to burial. Taking into consideration extensive development of cementing soil carbonate, the lithification of the Dashkovka paleosol appear to be pedogenic. Thus, Bed 46 is a calcrete crust or petrocalcic horizon in FAO classification (1998), and the Dashkovka profile in Zaborje represents durisol. The upper part of the calcrete that lacks primary sedimentary grains might have formed through calcification of the overlying soil horizon that was probably enriched in humus and corresponded to A horizon. The upper platy part of Bed 46 may represent a calcrete platy horizon. Platy horizons often form above impermeable hardpans (Wright, 1994). The top of Bed 46 is relatively smooth, which is untypical of calcrete tops and can probably be explained by transgressive erosion.

The Pre-Moscovian Paleosol

The upper limestone part of the section is superimposed by the mature subaerial exposure profile formed mostly during the Bashkirian Age until the Moscovian transgression. N. N. Bogolyubov was the first who noted in 1904 presence of this hiatus in the Kaluga Region (Khimenkov, 1927, pp. 33–35). Khimenkov (1927) and Shvetsov (1938) provided descriptions of this unconformity, respectively, in the northwestern and southern parts of the Moscow Syneclise. In Zaborje, the pre-Moscovian profile has typical appearance. The karst features in beds 43–54 include horizontal and vertical fissures, cavities, megavugs, karstic sinkholes, and collapse breccia. The mottled variegated clay partly or completely fills up karst cavities. The walls of cavities

Plate I. The Muratovka Paleokarst and microfacies of the uppermost Steshevian.

(1–5) Muratovka Paleokarst: (1) the field photo: (2a) cemented limestone, (2b) breccia grading into powdery horizon, (2c) dark brown clay at the profile top; (R) branching dissolution channels formed after root channels; (vn/tr) the Venevian/Tarusian boundary; note that limestone blocks in the base of Tarusian are displaced in the fractured zone (T); the objective cap is 65 mm in diameter. (2–5) signs of vadose diagenesis in thin sections: (2) brachiopod–ostracod–foraminiferal wackestone partly replaced by vadose micritic laminate, 1.1 m below the profile top, (3) rootlet voids in the primary packstone of the top of 2a; note that the voids are filled with geopetal vadose sediment, (4, 5) a limestone nodule from 2b: (4) secondary cavities with micritic halos, probably root cavities filled with geopetal sediment vadose silt and sparite cement, primary packstone is strongly recrystallized, (5) a crystalline veinlet rimmed by compact micrite and filled with characteristic vadose cement: earlier notched and later equant sparite; (6, 7) microfacies of the uppermost Steshevian Beds below the Dashkovka paleosol: (6) Bed 45, oncoidal grainstone–rudstone with alternating fine- and coarse-grained laminae, the rock is tightly cemented by equant sparite, (6) Bed 43, compacted argillaceous and silicified ostracod–oncoidal limestone. Bars are 1 mm in Figs. 2–4, 6, 7, and 0.5 mm in Fig. 5.

Plate II

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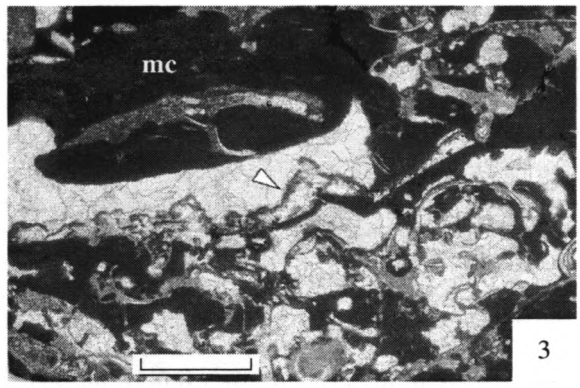

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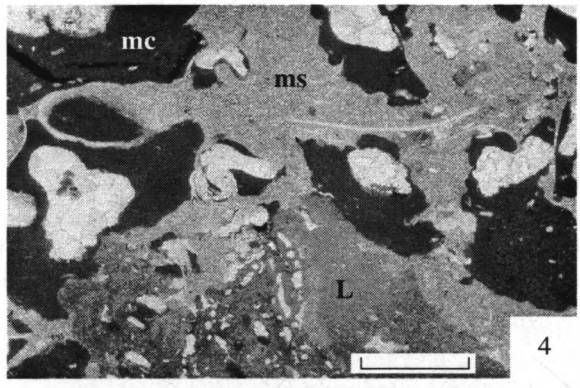

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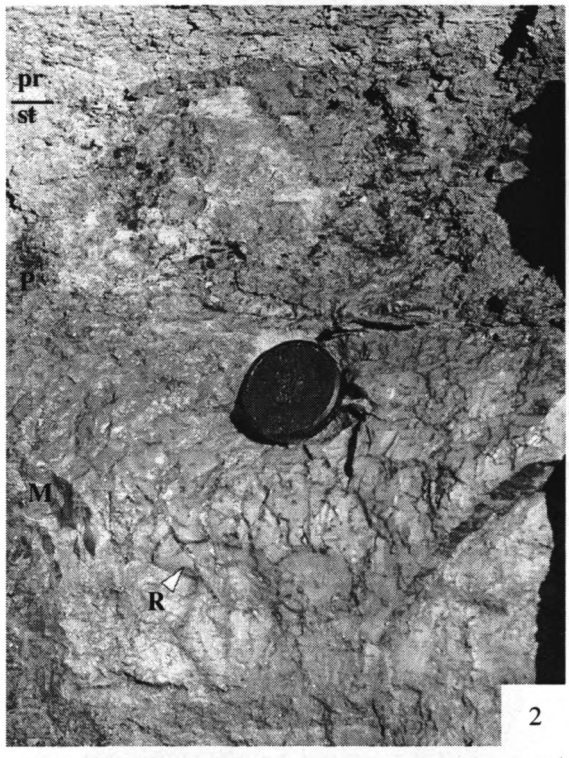

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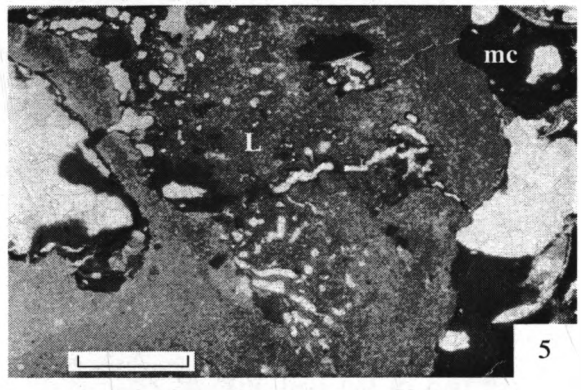

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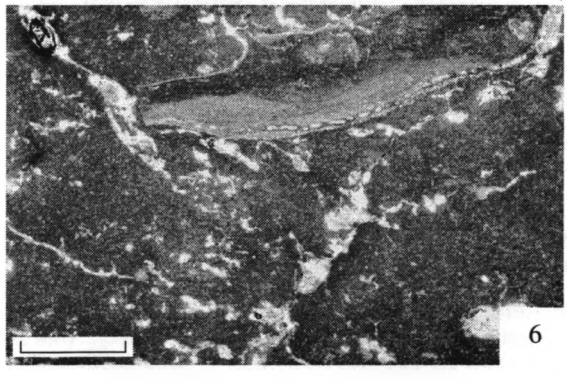
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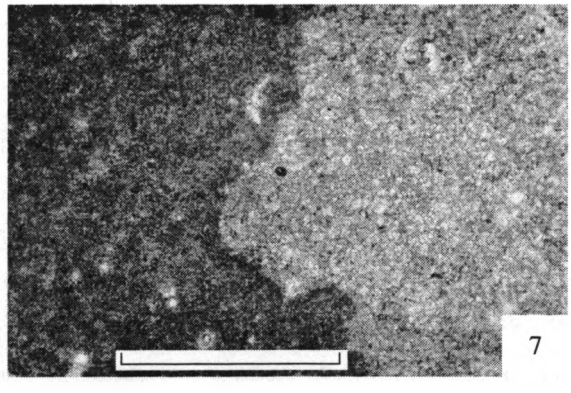
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are lined with laminar clay coatings. Some horizons within the limestone succession are intensely stylolitized. The top of the Protvian limestone along the eastern quarry wall shows the relief of 3 to 4 m. The rock in depressions is converted into breccia. The table-shaped elevations between depressions retain the original bedding of the limestone. The Protvian part of the section was described in one of these elevations (Kabanov, 2003). The limestone is strongly and unevenly silicified and recrystallized up to "sugar-like" or marble state. The breccia in upper horizons of the profile almost entirely consists of chertified fragments. Thin sections show extensive recrystallization and many generations of dissolution vugs filled by equant sparite cement. Later generations of vugs may remain uncemented, and some of them are filled by clay. The Vysokoe clays overlying and penetrating the karstified limestone are variegated, with alternated white, pale green, bright red, and ochric mottles. Red colors are moderately dominating. Shvetsov (1938) indicates that the SiO_2 content is 28.49–44.11% near the village of Vysokoe south of Serpukhov and 24.7–47.10% in Mikhailov Town; the Vysokoe clays are enriched in Al_2O_3 more than twice in comparison to lower Serpukhovian shales. Shvetsov (1938, p. 96) wrote that "the enrichment in Al_2O_3 both in the form of kaoline and free hydrates...suggests that this land environment...was favorable for the laterite or, at least, terra rossa weathering, i.e., the continental climate was hot enough, with periodical wetting". Kuznetsova *et al.* (in press) have revealed the smectite composition and absence of kaoline in the Vysokoe clay in Dashkovka, which do not permit attribution of this paleosol to terra rossa. The Al_2O_3, SiO_2, Fe_2O_3 and MgO content decrease down the profile, which is suggestive of weakly flushing regime, when these elements do not reach lower soil horizons. Preserved fragments of source limestone along the whole profile and absence of kaoline may suggest relatively weak weathering. Prevalence of red coloration over gley and presence of secondary carbonate nodules indicate moisture deficiency during soil formation, which may point to semi-arid environments.

MOTTLING IN UPPER VISAN AND LOWER SERPUKHOVIAN LIMESTONES

Tarusian and lower Steshevian limestones and marls feature dark gray cementation mottling. The mottles are harder than surrounding rock and confined to middle parts of limestone beds. Cementation mottles sometimes follow bioturbation traces. Shvetsov (1938; Shvetsov and Birina, 1935) recognized hard dark limestones of the type I, lighter soft fissile limestones of the type II, and nodular limestones of the type III where hard and soft varieties are alternating. In the Tarusian, the mottles are have very dispersed boundaries and are often indistinguishable in the type I hard cemented beds that are separated from each other by softer more or less fissile interbeds of the type II. In thin sections, the difference between types I and II is that hard limestone is cemented by and partly recrystallized to equant sparite. The equant sparite cement that coarsens toward pore centers is a typical product of near-surface freshwater diagenesis (Longman, 1980). Lack of early compaction features (Kabanov, 2003; Plate I, fig. 2) restricts cementation to early diagenetic stages. Alternation of compacted and uncompacted limestones, or limestone-marl alternations, is widespread in platform carbonate successions (Bathurst, 1987).

The type III limestones are characteristic of upper Visan limestones and especially distinct in the Venevian (Shvetsov and Birina, 1935). The nodular limestones were interpreted as "syngenetic marine breccia" (Shvetsov, 1938) and later called "intraclastic breccias" (Osipova and Belskaya, 1975; Makhlina *et al.*, 1993). Extensive intraclast formation could not be explained without excessive imagination, for instance sediment breaking by seismic sea-quakes (Vishnyakov, 1962; Osipova and Belskaya, 1967). However, the nodules show transitional zones to the surrounding softer rock and are tightly cemented by equant sparite (Shvetsov and Birina, 1935, plate V, fig. 22; plate VI, fig. 25). The soft rock adjacent to nodules is characterized by "fluidality" (ibid, Plate VI, fig. 26), which was subsequently called microlamination (Osipova and Belskaya, 1967; Osipova *et al.*, 1972; Makhlina *et al.*, 1993). The fluidality of Shvetsov appears to be the compactional fissility. The limestones of types I and II are usually thoroughly bioturbated, and re-orientation of bioclasts in a

Plate II. Lithofacies of the uppermost Steshevian and Dashkovka paleosol.

(1) The base of Bed 45, the ostracod–oncoidal facies, note the eroded top of micritic limestone covered by stromatolithic "flakes" and overlain by grainstone. (2–7) the Dashkovka paleosol: (2) field photograph: (M) massive calcrete, (P) platy calcrete, arrow indicates root impressions (R), (st/pr) top of the paleosol coinciding with the boundary between the Steshevian and Protva beds; it is poorly seen in the observation point; the objective cap is 65 mm. (3–5) base of the Dashkovka calcrete (Bed 46), highly calcretized molluscan grainstone–rudstone; the arrow in fig. 3 indicates lacy texture with large micritic calcans (MC), superimposed microsparite mottles (MS), and the section of a large rhizolith (L) composed of denser microsparite; small tubes piercing the rhizolith must have been left by rootlets; note sinusoidal boundaries between calcans and pores; cavities within calcrete are filled with equant sparite. (6) top of the Dashkovka calcrete, a dismicrite with presumable plant remains and small fissure-like fenestrae. (7) a contact between micritic phase of calcrete and microsparite filling of the root tube; sample is taken under the objective cap in Fig. 2. Bars are 1 mm in figs. 1, 3, 4–6 and 0.5 mm in fig. 5.

fluidal manner is a consequence of mechanical compaction.

The difference in the allochem composition between nodules and surrounding soft limestone has been put forward as evidence of "intraclast" migration (Shvetsov and Birina, 1935). However, this difference is easily explained by selective corrosion of micritic components in the soft limestone during compaction. The lower Venevian nodular limestones in the Novogurovskii section show that nodules have never been rounded, covered by micritic crusts, or bored, as may be expected for intraclasts. These potato-shaped nodules are sometimes penetrated by large tubular cavities up to 1 cm across, similar to the nodules in the powdery horizon of the Muratovka paleokarst. The Mikhailovian–Venevian succession of the southern Moscow Syneclise contains several paleokarst profiles, and nodules in the underlying limestone units appear to have formed through dissolution-reprecipitation of the carbonate sediment in a vadose setting. Many, if not all, nodules may be the cementation sheaths around roots. The origin of mottling in the lower Serpukhovian is less clear because appreciable subaerial exposure profiles are lacking. If cementation and recrystallization leading to mottle and hard bed formation is meteoric, vadose fluids might have been introduced into sediments during shoalings that correspond to formation of rooted horizons. However, the volume of aggressive meteoric water supplied by this way would hardly be sufficient for pervasive diagenetic changes. Rather, internal diagenetic processes such as aragonite dissolution and reprecipitation as calcite might have been involved.

CONCLUSIONS

(1) Two large cyclites (seventh-order rhythms of Makhlina *et al.*, 1993) are recognized in the Serpukhovian part of the section. The Tarusian-Steshevian interval correspond to the lower Serpukhovian cyclite, and the Protvian part of the section belongs to the upper Serpukhovian cyclite. Shoaling levels with rooted horizons mark the boundaries of lower-order cyclites. The Tarusian consists of two small cyclites and the Steshevian of two or three ones. The Protvian part of the section comprises at least one smaller-scale cyclite.

(2) Three subaerial exposure profiles are recognized. The base of the section corresponds to the *Muratovka paleokarst* known previously as a "rhizoid limestone" at the top of Venevian. The Dashkovka Paleosol at the top of Steshevian is described for the first time. It is characterized by well-developed calcrete hardpan and corresponds to durisol in FAO classification. The *Pre-Moscovian subaerial complex* is expressed in karstic terrane upon uppermost Steshevian–lower Protvian limestones and the Vysokoe clay drape. The profile is externally similar to terra rossa, however, the clay composition is predominantly smectitic.

(3) Nodularity of the upper Visan limestones is attributed to vadose alteration. Intraclastic interpretation of nodules is rejected. The origin of mottles and cemented beds in the lower Serpukhovian limestones and marls may similarly have subaerial origin.

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Some Unsolved Problems of Triassic Stratigraphy in Eastern Europe

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Abstract—Two problems of the continental Triassic stratigraphy of East Europe are considered: (1) correlation between the Lower Triassic horizons of the Moscow syneclise and the northern Caspian region and (2) correlation of the Middle Triassic horizons of the southern Cis-Urals and northern Caspian region with subdivisions of the International Stratigraphic Scale. The interdepartmental meeting on Triassic stratigraphy of the northern Caspian region (1999, Saratov) decided to substitute the former Ershov and Baskunchak regional horizons by the Lower Triassic Vetluga and Yaren superhorizons of the Moscow syneclise, respectively. However, the unification, according to which the Yaren age is accepted for deposits with Baskunchak charophytes in the Moscow syneclise, is not substantiated by occurrence of associated terrestrial fossil vertebrates. There is evidence that the Yaren Superhorizon of the stratotype area may differ in range from the corresponding interval in Caspian sections, since the latter probably includes analogs of the upper Vetluga Superhorizon. The boundary position between two Middle Triassic units of the southern Cis-Urals, i.e., between the Donguz Horizon with the *Eryosuchus* fauna of tetrapods and the Bukobai Horizon with the "*Mastodonsaurus*" fauna (found also in the upper Inder Horizon of the Caspian region), remains controversial. According to micropaleontological data, it coincides with the Anisian–Ladinian boundary, being placed inside the upper Ladinian based on vertebrate faunas. The reasons responsible for controversial viewpoints may be as follows: (1) the Germanic and Alpine Triassic section used as a reference by dating the East European paleontological remains can be imprecisely correlated; (2) the vertebrate-based correlations are based on ornamentation on dermal bone of amphibians (plagiosaurs), the evolution rate of which could be different in Central and Eastern Europe); and (3) of microfauna and palynological records can be of insufficient resolution ability for the long-distance correlations.

Key words: Triassic, stages, horizons, correlation, terrestrial vertebrates, palynology, ostracods.

The latest scheme of Triassic stratigraphy in the Caspian syneclise, which was adopted in 1999 (*Resolution...*, 2001), is essentially different from the previous one (*Resolutions...*, 1982). However, the decision did not unambiguously solve the basic problems of Triassic stratification in Eastern Europe as a whole. Some of these problems are discussed below.

LOWER TRIASSIC

The most significant problem is correlation of the Lower Triassic deposits of the Caspian syneclise with their northerly counterparts in the Moscow syneclise and the Volga–Ural anteclise (Figs. 1–3, Table 1). Such a correlation is of prime significance for the Lower Triassic stratigraphy in the east of the East European platform. A.N. Mazarovich whose hypotheses contributed much to progressing of the continental Triassic stratigraphy suggested a simple solution of the problem, the facial persistence of corresponding deposits parallel to the Urals. Based on the idea, he distinguished the uniform formations extending them from the northern platform areas to the Mt. Bol'shoe Bogdo in the south. The principal subdivisions were the Buzuluk Formation included into the Vetluga regional stage and the Tananyk Formation eventually attributed to the Baskunchak regional stage of the Caspian region together with the overlying Bogdo Formation of marine facies.

Since the 1950s, when the drilling program was in progress, the new data showed a more complex structure of the Lower Triassic, and the formations after Mazarovich were abandoned. Blom and Ignat'ev (1955) suggested a new rhythmostratigraphic approach to deposits of the Moscow syneclise. Based on paleontological evidence (finds of *Ceratodus*), Blom recognized here the upper sedimentary rhythm (Fedorovka Horizon) correlative the Baskunchak strata. Later on, it was established however that rhythms, which were regarded as readily recognizable even without a sufficient paleontological characterization, appear to be unsuitable for a reliable correlation. For this reason, Blom failed to elaborate in detail a well-substantiated stratigraphic scheme. This was an

†Deceased.

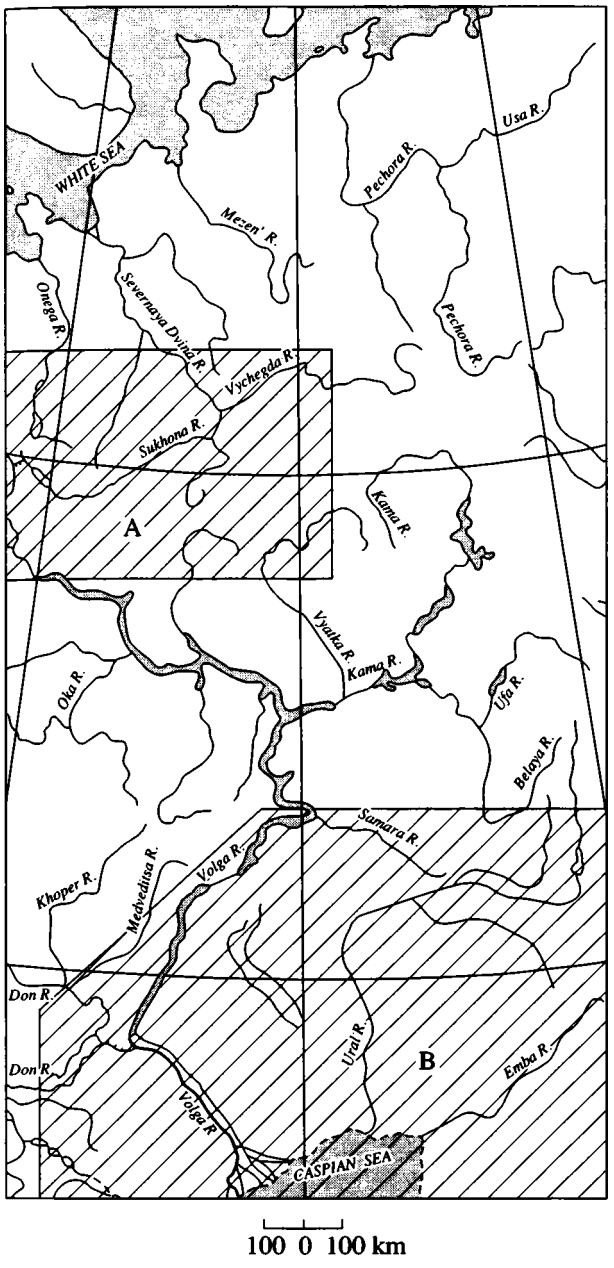


Fig. 1. Geographic areas of Triassic sections studied in Eastern Europe: (A) Moscow syneclise; (B) Cis-Urals and Caspian syneclise.

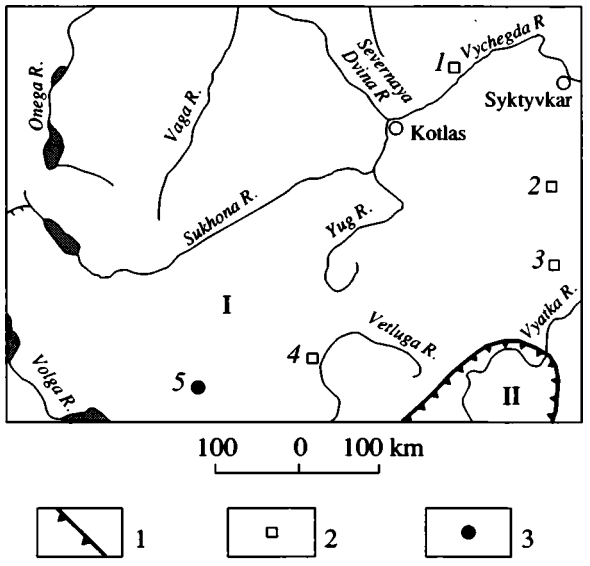


Fig. 2. Lower Triassic sections in the Moscow syneclise: (1) boundary of structural area; (2) natural exposure; (3) borehole; (I) Moscow syneclise; (II) Volga–Ural anteclise. Exposures and boreholes by sites: (*1*) Lopatino II, (*2*) Yagshordyn, (*3*) Fedorovka, (*4*) Zubovskoe, (*5*) Galich depression.

important lesson to remember. Nevertheless, the rhythmostratigraphic approach stimulated a comprehensive study of fossils distribution in the Lower Triassic sequences, especially of tetrapods, the most informative among the others.

The modern scheme of the Lower Triassic stratigraphy in the Moscow syneclise is based on terrestrial vertebrates. Being connected with early works by Efremov, it was elaborated by Shishkin who studied tetrapods and by Lozovsky who used the geological survey data to control of the succession of assemblages. As it turned out, the scheme based on all the data on terrestrial vertebrates (Shishkin and Ochev, 1967), fish remains (Minikh, 1977), and results of rhythmostratigraphy (obtained by Strok, Tverdokhlebov, and other researchers) is applicable in a vast area extending southward up to the Obshchii Syrt and southern Cis-Urals. With later corrections (Lozovsky *et al.*, 1968, 1991; Novikov *et al.*, 1990), it is possible to recognize here the Vokhma, Rybinsk, Sludka, and Ust'-Myla (insufficiently studied so far) horizons of the Vetluga Superhorizon and the Fedorovka and Gam horizons of the Yaren Superhorizon (Table 1).

In the northern Caspian region, where Triassic deposits penetrated by drilling are enormously thick (Figs. 1 and 3), their stratification is mainly based on micropaleontological records but not on tetrapod fauna, thus representing "a peculiar realm" as compared to stratigraphy of coeval deposits in northern areas of the platform. After recognition of thick Upper Triassic sequences, V.V. Lipatova and her collaborators obtained new important data. They proved a wide distribution of Middle Triassic marine deposits and worked out a new stratigraphic scheme of the Lower Triassic divided into the Ershov and Baskunchak horizons, which was approved in 1979 (*Resolutions...*, 1982). The Baskunchak Horizon stratotype corresponds to the synonymous group distinguished in the southwestern part of the syneclise, where it is divided (from the base upward) into the Akhtuba, Bogdo, and

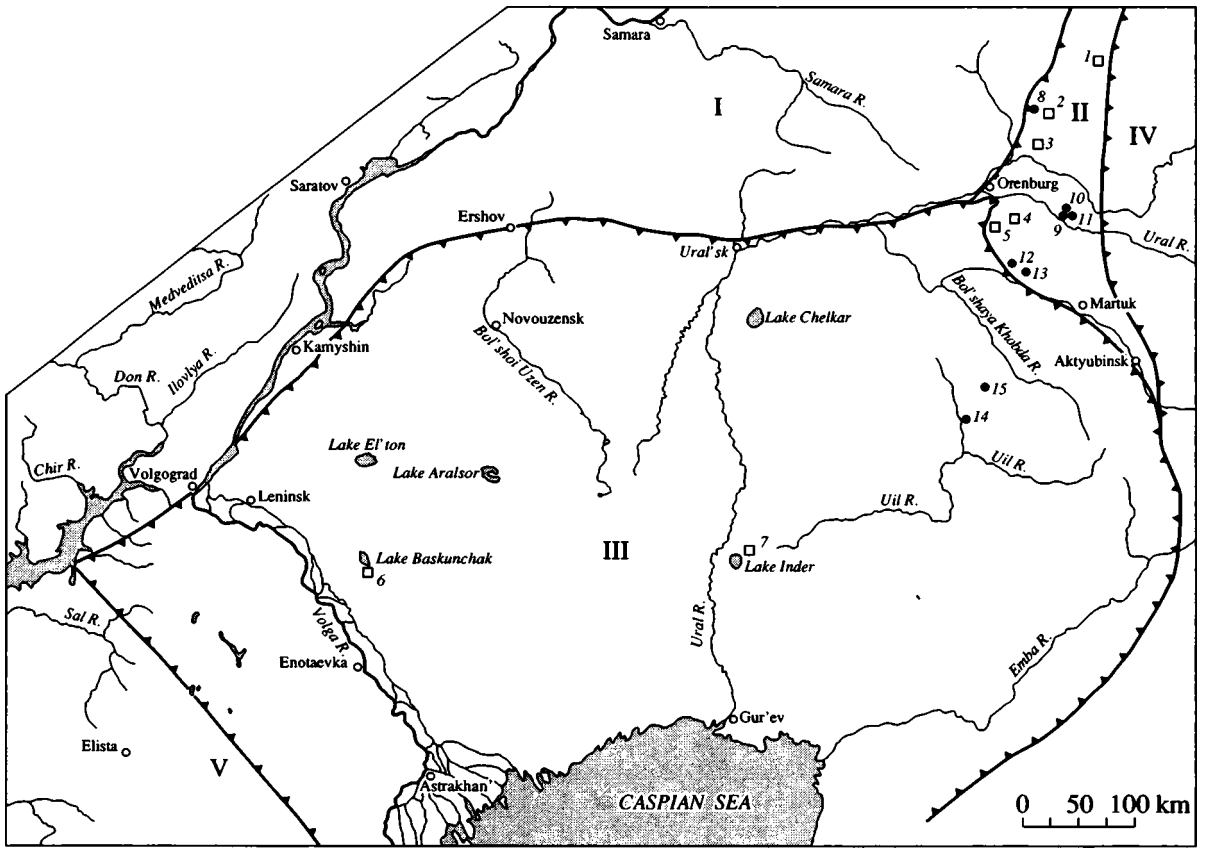


Fig. 3. Main Triassic sections in the Uralian foredeep and Caspian syneclise: (I) southeastern slope of the Volga–Ural anteclise; (II) Uralian foredeep; (III) Caspian syneclise; (IV) fold system of the Urals; (V) Karpinsky rampart; (*1*) Surakai, (*2*) Staroe Koltaevo, (*3*) Petropavlovka, (*4*) Bukobai, (*5*) Donguz, (*6*) Bol'shoe Bogdo and (*7*) Kara-Bala-Kantemir naturally exposed sections; (*8*) Borehole 150, Staroe Koltaevo site; (*9*) boreholes C-36, (*10*) 15 and (*11*) 17 in Novocherkassk basin; (*12*) Borehole 11-C, Tamarkuduk basin; (*13*) Borehole 5, Kairat site; (*14*) Khobda reference borehole; (*15*) borehole on Krasnoyar Dome (symbols as in Fig. 2).

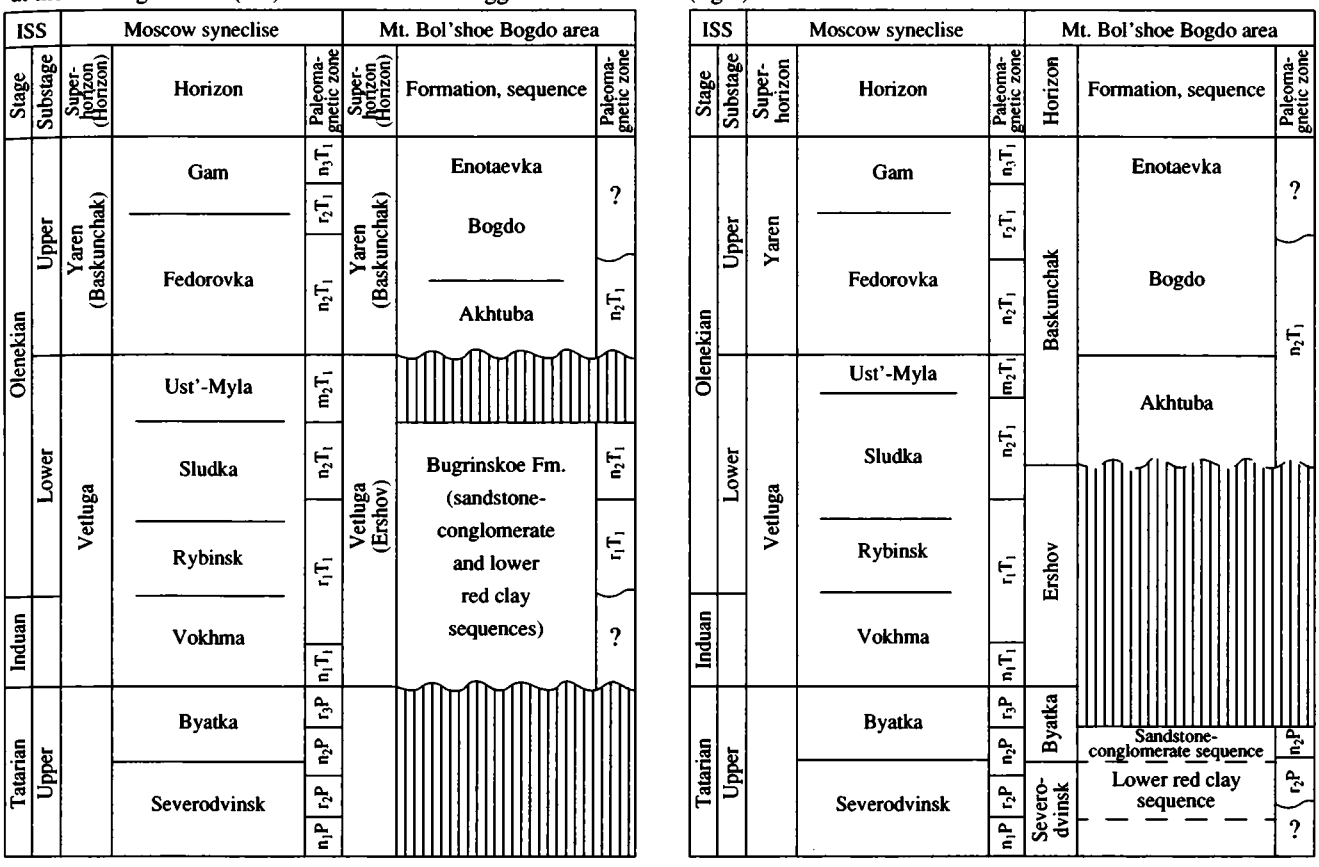
Enotaevka formations (Table 1).[1] The Ershov and Baskunchak horizons were discriminated based on distribution of charophytes and correspond to charophyte zones I–II after Saidakovskii (zone I after Kiselevskiĭ) and zone III after Saidakovskii (zone II after Kiselevskii), respectively.

Before the meeting of 1979, the Lower Triassic was subdivided into the Vetluga and Baskunchak groups (stages of Mazarovich) in the entire eastern part of the platform. To the north of the Caspian syneclise, the boundary between groups separates, in terms of present-day nomenclature, the Ust'-Myla Horizon with its *Vetlugosaurus* fauna of tetrapods from the Fedorovka Horizon bearing the *Parotosuchus* fauna. In the Caspian region, this boundary was placed at the base of the Akhtuba Formation (Tananyk Formation of the Bol'shoe Bogdo section, according to Mazarovich). The horizons are mainly discriminated based on the mentioned charophyte zones. Representatives of the *Parotosuchus* tetrapod fauna characteristic of the northerly Fedorovka and Gam horizons are known only from the Bogdo Formation of the Caspian region, where underlying Triassic deposits yield no vertebrate remains. Accordingly, it is impossible to trace the boundary from northern to southern sections based on these poorly correlating paleontological records.

In accordance with recommendations of "The Stratigraphic Code of the USSR", the meeting of 1979 ranked former groups as regional horizons. Because the Baskunchak charophytes were detected at that time in the upper Vetluga deposits of the Moscow syneclise, the meeting discredited the unified horizons for the entire eastern part of the platform. In the north, where the Vetluga deposits were ranked as synonymous superhorizon, there was discriminated the Yaren Horizon (now a superhorizon; Lozovsky *et al.*, 1991) corresponding

[1] In the regional stratigraphic scheme of 1999, these formations are ranked as horizons, although the original subdivisions are untraceable outside the Caspian stratotype area. The horizons are incorporated into the Yaren Superhorizon, which includes the other horizons northward of the Caspian region. The situation like this is inconsistent with regulations of "The Stratigraphic Code" (1992), which unambiguously forbid a possibility to divide horizons into subhorizons, as it would be difficult to control synchronism of horizons and superhorizons over the area, where they are recognizable.

Table 1. Variants of correlation between Lower Triassic sections of the Moscow syneclise and Mt. Bol'shoe Bogdo area; one accepted at the meeting of 1999 (left) and the other one suggested in this work (right).



M/z—magnetic polarity zones.

in range to the upper part of the Baskunchak Horizon of the Caspian region. The deposits below the Baskunchak Horizon, which were previously correlated with the entire Vetluga interval, were attributed based on charophytes to the Ershov Horizon of the Caspian region, which corresponds only to the Vokhma Horizon, the lower one in the Vetluga Superhorizon. Position of the Vetluga–Yaren boundary remained indefinite therewith in the south.

The stratigraphic position of beds with the Baskunchak charophyte assemblage (zone II after Kiselevskii) remains disputable in the Moscow syneclise. The assemblage is recovered not from the upper reference subdivisions of the Vetluga Superhorizon (Rybinsk and Sludka horizons), which yield tetrapod remains, but from structurally correlative sequence recovered by drilling in the Galich depression (Fig. 2) (Strok *et al.*, 1984). The correlation was doubted later on, and the sequence was attributed to the Pervushino Formation of the Yaren Age based on comparable cherry–red coloration of the Pervushino and Fedorovka clay beds (Lozovsky, 1992).

The new age interpretation of charophyte-bearing deposits of the Moscow syneclise persuaded the meeting of 1999 to return to the Lower Triassic stratigraphic scheme used until 1979 in the eastern part of the platform, in which the Akhtuba Formation of the Caspian region was included into the post-Vetluga interval (Table 1, left side). As a consequence, the meeting suggested to regard the Vetluga and Yaren superhorizons as units traceable throughout the region, which may substitute the Ershov and Baskunchak horizons, respectively, in the Caspian region. In our opinion, this decision is far from being unambiguous.

As noted above, the decision of 1999 (*Resolution...*, 2001) is based on the revised age of deposits with Baskunchak charophytes in the Moscow syneclise. Actually, there was no other reason for revision except for similarity between these deposits and clays of the Fedorovka Horizon. Such a diagnostic criterion is of little significance for Lower Triassic deposits of Eastern Europe. Moreover, the incorporation of the Caspian Akhtuba Formation into the Yaren Horizon is in disagreement with the following facts.

(1) The transitional assemblage of dipnoan *Gnathorhiza–Ceratodus* characteristic of the lower Yaren Horizon in northern areas (the Fedorovka, Lopatino I, Lopatino II, and Yagshordyn localities of Fedorovka age; Fig. 2) occurs high in the Baskunchak Horizon:

about 5 m below the top of the Akhtuba Formation that is not less than 55 m thick.

(2) Labyrinthodont species *Parotosuchus bogdoanus* (Woodw.) known from the lower Bogdo Formation (Mt. Bol'shoe Bogdo locality) is closer to forms of the *P. orientalis* (Otschev) group occurring in the lowermost Yaren Horizon than to those of *P. orenburgensis* (Konzh.) group found higher in this horizon. These forms differ from each other in morphology of tabulare "horns," which are thicker and curved in early *Parotosuchus* forms, such as *P. orientalis* and *P. helgolandicus* (Schroeder) or non-described species from the Fedorovka Horizon section at the Fedorovka and Luza rivers, whereas younger *O. orenburgensis* and *P. nasutus* (Meyer) had flattened and straight horns. The right tabulare base preserved in a skull fragment of *P. bogdoanus* (collection of PIN, no. 2242/23) from the Bogdo Formation is thickened like in ancestral forms.

(3) The Bogdo Formation can be correlated based on ammonites with the lower part of the upper Olenekian (the *Tirolites harti* Zone, Lozovsky, 1992), i.e., with the Yaren Horizon base.

(4) Finally, the ostracod assemblage similar, in opinion of Kukhtinov, to the Baskunchak one was discovered by Starozhilova from the upper Vetluga Superhorizon of the Cis-Urals (the Kzyl-Sai Formation section near the village of Petropavlovka, the Orenburg oblast; Fig. 3). The assemblage includes *Darvinula postparallela* Mish., *D. sedecentis* Mand., *D. designata* Schn., *D. acuta* Mish., *D.* aff. *acmayica* Schl., *D.? dubia* Starozh., *Suchonella* aff. *buginella* Mish., *Darwinula oblonga* Schn., *D. parva* Schn., *D. aceris* Mish., *D. obliqua* Gleb., *D. temporalis* Mish., *Gerdalia clara* Mish., and other forms (Tverdokhlebov, 1967). It should be noted that the correlation between the Caspian and Moscow Triassic based on ostracods is ambiguous in general, although the above ostracod assemblage is from a bordering area that was likely connected with the Caspian region in terms of zoogeography. The ostracod-bearing bed is situated 25 m below the base of the Yaren Horizon (the stratotype section of the Petropavlovka Formation with remains of labyrinthodonts Trematosauridae at the base). The overall thickness of the Kzyl-Sai Formation enclosing remains of Vetluga labyrinthodont *Wetlugasaurus* is 114 m. Accordingly, the Baskunchak ostracod assemblage from the Caspian region may characterize both the Yaren and post-Vetluga times.

Data of independent methods (in this case paleomagnetic) may clarify ambiguous relationship between the Vetluga–Yaren and Ershov–Baskunchak boundaries. Paleomagnetic measurements have been carried out in the mentioned section of the Mt. Bol'shoe Bogdo near the Lake Baskunchak (Table 1). From the top downward there are exposed the fossiliferous lower Bogdo and Akhtuba formations and barren sandstone–conglomerate and lower red clay sequences commonly attributed to the Bugrinskoe Formation, an equivalent of the Ershov Formation distinguished in the northwest of the region. The boundary between the Akhtuba Formation and sandstone–conglomerate sequence is distinctly unconformable. It is exactly this boundary that was considered as corresponding to any of two significant biostratigraphic levels of the Lower Triassic of the Moscow syneclise, i.e., to either the Vokhma–Rybinsk (the meeting of 1979) or Vetluga–Yaren (the meeting of 1999) boundaries.

In the last case, we get a simple interpretation of paleomagnetic zones r_1T_1 and n_2T_1 established in the lower Sludka and upper Sludka–Fedorovka intervals of the section (Table 1, left side). However, the resolution of 1979 also admits a plausible interpretation that is connected with idea of Mazarovich who argued for occurrence of the Upper Permian red clays in the Bol'shoe Bogdo section, which are attributed now to the Bugrinskoe Formation. In addition, the conclusion of Rykov (1958) that the boundary between the sandstone–conglomerate sequence and overlying red-clay unit (now the Akhtuba Formation) corresponds to the Permian–Triassic boundary, deserves attention. In such a case, there may be a large hiatus spanning the uppermost Permian–lowermost Triassic interval. The Akhtuba interval in the normal polarity zone can be correlated then with n_2T_1 zone in the Sludka Formation sequence, whereas sandstone–conglomerate beds spanning the lower part of the normal polarity zone can be regarded as analogs of the Vyatka deposits within paleomagnetic zone n_2P (Table 1, right side). Correlation of the zone lower part with the interval n_2T_1 would be inconsistent with the fact that any noticeable unconformity corresponding to the Permian–Triassic boundary and recognizable throughout Eastern Europe is missing below. The occurrence of the upper Tatarian deposits on salt stocks and incompleteness of their Vyatka part are not the extraordinary facts in the western Caspian region. According to Kukhtinov, the relationships like these are inferable from microfaunal records.[2]

Accepting the above interpretation, we meet another contradiction, i.e., the absence of Subzone 2 in the interval n_2T_1, which should correspond to the Ust'-Myla Horizon of the upper Vetluga Superhorizon. In the correlation chart approved at the meeting of 1999, the subzone absence may be to the Lower Triassic hiatus between the Akhtuba Formation and the sandstone–conglomerate sequence belonging to the upper Bugrinskoe Formation. However, this subzone (which has not been detected in any continuous paleomagnetic records of the East European Triassic) may be discovered after additional detailed investigations. Because of the low

[2] The absence of Permian deposits in the Bol'shoe Bogdo exposures was suggested based on drilling data implying an enormous thickness of the Lower Triassic in the northern Caspian region (Ochev, 1967). The suggestion has not been proved by facts however. The absence of Permian deposits on highly elevated domes in the western part of the depression (Lozovsky *et al.*, 1973) is also of a little significance because of disputable position of the Permian–Triassic boundary in many sections.

sedimentation rate during the Akhtuba time, this subzone may be of a minor thickness. In general, the paleomagnetic scale of Eastern Europe is extremely generalized and lacking many polarity subzones established in marine sections. Comparing it with the scale recommended in "The Stratigraphic Code of Russia" (*Supplement...*, 2000), one easily see that many polarity reversals are missing.

The above review is aimed not to argue for one of the alternatives (Table 1) but to emphasize the impossibility to make a choice at present. In other words, the accepted correlation scheme for the Lower Triassic in Caspian and northern regions, which is based on unified superhorizons, creates an impression of clearness that has not been gained in fact. The problem can be solved if stratigraphically important microfossils (charophytes) will be found in association with informative vertebrate remains of the upper Vetluga Group. The large tetrapod burials, for example, that of *Wetlugasaurus angustifrons* Riab. in the Sludka Horizon near the village of Zubovskoe at the Vetluga River, are the best localities to seek for the microfossils. The indicated section includes thin interbeds of bluish gray clays (Efremov and V'yushkov, 1955) perspective for microfossil research. The mentioned Petropavlovka section of the southern Cis-Urals is also of a great potential.

The reliable Triassic correlations need also in revision of the paleomagnetic scale. The scale available at present includes many subzones of different ranks hardly usable in practice, and this created a tendency to generalize the real patterns at the expense of details. For example, prior to the meeting of 1979, the normal polarity subzones of zone r_1T_1 were established by Molostovskii and Tverdokhlebov in the Obshchii Syrt section (Tverdokhlebov, 1975) and by Ochev (1987) in the Vokhma Horizon exposures along the Vetluga River. The subzones have not been included however in the paleomagnetic scale adopted at the meeting. The tendency to generalize the scale disagrees with interests of experts in bio- and lithostratigraphy who need subzones for detailed correlation. The detailed paleomagnetic records have been successfully by correlation of the Lower Triassic deposits in the Obshchii Syrt and southern Cis-Urals and by substantiation of the Ust'-Myla Horizon.

MIDDLE TRIASSIC

As for the Middle Triassic deposits in European Russia, which are widespread in the Cis-Urals and Caspian syneclise (Figs. 1, 3; Table 2), the problematic issue concerns their real stratigraphic range and correlation with subdivisions of the International stratigraphic scale (ISS). The problem of prime significance is position of the Anisian–Ladinian boundary in the regional reference sections. In the Lower Triassic correlation scheme substantiated by Shishkin and Lozovsky, the Vokhma Horizon corresponds to the Induan, the Rybinsk and Sludka horizons to the lower Olenekian, and the Yaren Superhorizon, to the upper Olenekian that is accepted widely. On the contrary, correlation of the Middle Triassic with the ISS is ambiguous, meeting serious contradictions, which were discussed in several publications (Kukhtinov, 1999; Ochev, 1999; Shishkin and Ochev, 2002). The main way to eliminate the uncertainties is a thorough study of highly fossiliferous Middle Triassic sequences widespread in the southern Cis-Urals and northern Caspian region. Their stratigraphic intervals (Table 2) correspond to the Donguz and Bukobai formations in the former region and to the El'ton, Inder, Masteksai, and Akmamyk formations and their analogs in the latter region (nomenclature after the unified regional stratigraphic schemes; Kukhtinov *et al.*, 2001). The long-distance correlations are mainly based on palynological and ostracods assemblages, and on remains of terrestrial vertebrates. Ages of subdivisions are usually inferred based on correlation with the Triassic deposits in the Germanic basin and, directly or indirectly, with the Alpine scale.

In general, there are three variants of correlation and dating the Middle Triassic deposits in Eastern Europe: (1) the basic variant shared by majority of experts in micropaleontology (Tuzhikova, 1975, 1979; Makarova, 1975; Makarova and Vergaĭ, 1995; Kukhtinov, 1999; Il'ina, 2001; Yaroshenko *et al.*, 2001), (2) the variant based on tetrapod distribution (Ochev and Shishkin, 1989; Shishkin and Ochev, 1992, 1999), and (3) an essentially different variant of Movshovich (Movshovich and Kozur, 1975; Movshovich, 1980, 1998), which is based on ostracod records interpreted by Kozur (Table 2). Some principal points of the variants and their substantiation are analyzed below in order to reveal possible reasons of the discrepancies. We use therewith the correlation of Triassic deposits of the Germanic basin with the ISS, as proposed by Kozur (1974, 1998).

In the Caspian region, the most significant point of controversy is positioning of the Inder–Masteksai boundary, as it is defined based on different microfossils. According to the ostracod data, this boundary is inside the basal lower Ladinian (Fassanian), and the *Pulviella aralsorica* Zone, the upper one of two Inder zones, can be attributed to the upper Anisian–lower Ladinian interval (Kukhtinov (1999). This version is accepted by all Russian palynologists who studied the northern Caspian region. For example, the spore–pollen subassemblage PA-III (1) with *Microcachryidites–Distalanulisporites* from the upper Kiil Subformation of the reference Khobda Borehole section (the age analog of the upper Inder Formation in the eastern Caspian depression) ranges from the upper Anisian to the lower Ladinian (Yaroshenko *et al.*, 2001), but it may be confined to the upper Anisian only, as suggested by Il'ina (2001).

Formerly, it was also suggested, based on pelecypod and charophyte distribution, that the Inder Horizon corresponds to the Ladinian. The recent revision (Zhidovi-

Table 2. Alternative correlations between the ISS units and Middle Triassic subdivisions in the South Caspian syneclise and Cis-Urals region

ISS		Movshovich, 1998		Shishkin and Ochev, 1999		Kukhtinov *et al.*, 2001	
Stage	Substage	Southern Cis-Urals	Northern Caspian region	Southern Cis-Urals	Northern Caspian region	Southern Cis-Urals	Northern Caspian region
Ladinian	Upper	Surakai Fm.	Lower Khobda Fm.	Bukobai Fm.	Akmamyk and Masteksai Fms.	Bukobai Horizon	Akmamyk and Masteksai horizons
	Lower		Akmamyk and Masteksai Fms.	Donguz Fm.	Inder and El'ton Fms.		
Anisian	Upper	Bukobai Fm. / Yushatyr Fm.				Donguz Horizon	Inder and El'ton horizons
	Middle						
	Lower	Donguz Fm.	Kill (Inder) Fm.				

nov, 1998) showed however that the Inder pelecypods characterize only the Middle Triassic as a whole. Saidakovskii and Kiselevskii (1985) considered the charophyte assemblage with *Stellatochara hoellviciensis* and *Stenochara donetziana* (Zone IV spanning the Inder and Masteksai horizons) as comparable with charophytes from the upper *Ceratites* Beds and the lower Keuper of Germany, where these subdivisions are attributed to the upper Ladinian. Bilan (1988) established in Poland the analogs of the *Stellatochara hoellviciensis* Zone within the range of Upper Muschelkalk, which corresponds, according to Kozur (1998), to the greater part of Illyrian, Fassanian, and a part of Longobardian. Thus, the above charophyte assemblage implies the upper Anisian–upper Ladinian interval for host deposits, being unsuitable for a more accurate dating.

In the southern Cis-Urals (Fig. 3, Table 2), micropaleontological data indicate that the Donguz Horizon with the *Eryosuchus* tetrapod fauna is an approximate age analog of the El'ton and Inder horizons of the Caspian region. The basal part of the stratotype Donguz Formation, the eponymous river locality, is attributed based on palynological data to the middle–upper Anisian (Makarova and Vergaĭ, 1995; Il'ina, 2001; Yaroshenko *et al.*, 2001).[3] The upper part of the formation is dated in another section near the village of Staro-Koltaevo, southern Bashkiria, where deposits yield tetrapod remains. According to Tverdokhlebov who studied natural exposures of the same formation along the Bol'shoi Yushatyr River, the unit was recovered by Borehole 150 immediately below the basal Bukobai member (the stratotype of the Yushatyr' Formation after B.P. V'yushkov). At the level of 26 m below the top of the Donguz Formation (overall thickness 156 m), Makarova found the spore–pollen assemblage including the Alpine Illyrian (late Anisian) taxa *Distalanulisporites puncus* Klaus, *Concentricisporites nevesi* Antonescu, and *Triadispora crassa* Klaus, all known from the *Paraceratites trinodosus* Zone.[4]

The overlying Bukobai Formation characterizing the synonymous horizon was studied in the Bukobai Gully stratotype section near the village of Mikhailovka at the Berdyanka River (the left tributary of the Ural River). The formation encloses the *Mastodonsaurus* tetrapod fauna and rests unconformably on the erosional surface of the Donguz Formation. It is usually correlated with the Masteksai Horizon and now with the overlying Akmamyk Horizon of the Caspian region (Kukhtinov *et al.*, 2001). Specialists unanimously attributed the spore–pollen assemblages of the unit to the Ladinian Stage (Tuzhikova, 1975, 1979; Makarova, 1975; Makarova and Vergai, 1995; Yaroshenko *et al.*, 2001), whereas assemblages from the overlying gray deposits of the Surakai type are not older

[3] This correlation is substantiated in the text (Yaroshenko *et al.*, 2001) in disagreement with position of the *Paratrilites minor* assemblage shown in the table.

[4] The spore–pollen assemblage, which was previously attributed to the upper Donguz Formation and dated as Ladinian in age (Makarova and Vergai, 1995), was obtained not from the stratotype area but from the borehole the near the village of Kairat of the Martuk district, Aktyubinsk oblast. The enclosing deposits are referred now to the Donguz Formation conventionally, and the assemblage is defined as the Anisian in age (Yaroshenko *et al.*, 2001).

than the early Carnian, like their counterparts from the Surakai Formation stratotype in the eponymous basin.

It is still difficult to define the Bukobai Formation age more precisely. Some experts in palynology attribute this formation, completely (Makarova and Vergai, 1995) or without the basal beds (Tuzhikova, 1975, 1979), to the upper Ladinian. According to the other opinion (Il'ina, 2001), this subdivision spans the entire Ladinian. According to above publications, spore–pollen assemblages are uniform throughout the Bukobai Formation section and have been referred to the single of *Florinites pseudostriatus–Minutosaccus potoniei* local zone.[5]

At present, the thorough study of Middle Triassic spore–pollen assemblages from the Khobda reference borehole in the Caspian region clarified palynology of the Bukobai deposits (Yaroshenko *et al.*, 2001). In the Khobda section, the former assemblage of the Bukobai Formation is divided into the *Microcachryidites–Distalanulisporites* (partly), *Converrucosisporites conferte-ornatus–Illinites chitonoides*, and *Leschikisporites–Todisporites* spp. subassemblages, the second one attributed by researchers to the lower Ladinian. However, the palynological stratification that is necessary to specify the age range of Bukobai deposits has not been performed. Thus, the available data on microfossils suggest that the Inder–Masteksai and Donguz–Bukobai boundaries approximately correspond to the Anisian–Ladinian boundary.

Correlating the Middle Triassic deposits under consideration with concurrent sediments of the Germanic basin (Table 2) based on vertebrate remains, we arrive at different age assessments. The *Eryosuchus* and *"Mastodonsaurus"* faunas dominated by amphibians of the order Temnospondyli, which occur in the Donguz and Bukobai horizons respectively, are most important for stratigraphy of the southern Cis-Urals. The Germanic Middle Triassic includes three main intervals with non-marine tetrapods (predominantly amphibians of the same order). These are the Rhaetian (lower Anisian) *Eocyclotosaurus* fauna, which may include poorly studied fossils from the Lower Muschelkalk (basal middle Anisian), the Muschelkalk–Lettenkeuper boundary beds ("Grenzbonebed") with poorly studied amphibian remains, and the Lettenkeuper proper with the *Mastodonsaurus* assemblage. According to correlation charts by Kozur (1974, 1998), two latter intervals with tetrapod remains known mainly from Baden-Württemberg site correspond to the upper Ladinian.

Sections in the Cis-Ural lack analogs of the *Eocyclotosaurus* fauna and the early Anisian palynological assemblages. On the other hand, they host faunas similar to two younger Germanic assemblages in appearance and abundance of close amphibian-plagiosaur genera of the family Plagiosaurinae occurring nowhere else except Europe until the early Late Triassic. Successive species of these genera (*Plagiosuchus* of Central Europe and *Plagioscutum* of the Cis-Urals) demonstrate the same evolution pattern of dermal bone ornamentation from "vermiform" to "pustulate" (Shishkin, 1987). In the Lettenkeuper of the Germanic section, *Plagiosuchus pustuloglomeratus* Huene from Grenzbonebed[6] was succeeded by "pustulate" *P. pustuliferus* Fraas (Huene, 1922). According to this criterion, the *Eryosuchus* fauna from the Donguz Horizon of the Cis-Urals is correlative with fauna of the Muschelkalk upper horizon adjacent to Grenzbonebed, and the *Mastodonsaurus* fauna of the Bukobai Horizon is comparable with the Lettenkeuper tetrapod remains. In addition, both the Bukobai and Lettenkeuper tetrapod assemblages are dominated by giant amphibians mastodonsaurids, although the Uralian form described as *Mastodonsaurus* actually belongs to another close genus (Shishkin and Ochev, 1992). Since the Upper Muschelkalk and Lettenkeuper correspond to the upper Ladinian of the Alpine section, the boundary between two Middle Triassic horizons of the Cis-Urals can be defined based on the faunal evolution discussed above. The Bukobai Horizon with its tetrapod fauna should be not older than the upper Ladinian, and the upper limit of the Donguz Horizon is corresponding to the lower part of the upper Ladinian (Shishkin and Ochev, 1992).

Taking into account the present inadequate dating of horizons in the southern Cis-Urals (and a hiatus between them), ages estimated based on tetrapods do not differ much from inferences based on micropaleontological data. In particular, the late Anisian palynological assemblage of the upper Donguz Formation does not exclude a possibility that the *Eryosuchus* fauna might have its last occurrence in the Ladinian.

We can only state that burials of *Eryosuchus* and *Mastodonsaurus* faunas join each other at the boundary between the upper Donguz Formation (upper Anisian) and the Bukobai Formation (upper or unspecified Ladinian) that unconformably rests on erosional surface of the former. It may be possible that a large part of the Ladinian corresponds to the break in sedimentation and the renewal of fauna took place in the late Ladinian. In this case, the biostratigraphic boundary between corresponding horizons is at the same level.

In addition, plagiosaur remains of the Donguz type (*Plagioscutum ochevi?* Shishkin; Shishkin, 1987),

[5] Tuzhikova (1979) considered the palynological assemblage from the basal Bukobai sandstones (the former Yushatyr Formation) as older than that from the Bukobai Formation proper, which she attributed to the upper Ladinian. In contrast, Makarova believed that distinction between assemblages is a consequence of facies peculiarity of their host deposits (stream sediments), and that fluvial activity increased proportion of large forms with coarse ornamentation, which impart an "older" outlook to the Yushatyr assemblage.

[6] In the latest revision of the Germanic plagiosaurs (Hellrung, 2003), the holotype of *Plagiosuchus pustuloglomeratus* (a fragment of pectoral girdle) is attributed to *Gerrothorax*, but judging from the published photographs, its bad preservation gives little grounds for this interpretation.

which were found in association with ostracods of the *Lutkevichinella bruttanae* Schn. Zone and with the early–middle Anisian palynological assemblage (Makarova, 1975; Il'ina, 2001) in the El'ton Horizon of the Caspian region, can be indicative of the lower age limit of the *Eryosuchus* fauna.

At the same time, the absence of equivalent tetrapod fauna (plagiosaurs of the "vermiform" type) in the middle Anisian–lower Ladinian of the Germanic basin can be accounted for by the fact that this interval (Muschelkalk) is mainly composed of epicontinental sea facies unfavorable for preservation of terrestrial vertebrate remains (Shishkin *et al.*, 2000). All known "vermiform" plagiosaurines (attributed to *Plagiosuchus pustuloglomeratus*) from Thuringia and Lorraine of Central Europe and from Württemberg are confined to the Upper Muschelkalk of the upper Ladinian (Schoch and Werneburg, 1998).

In contrast to Middle Triassic subdivisions of the Cis-Urals, the microfossil-based ages of concurrent units in the Caspian region are in evident disagreement with inferences based on tetrapods. The main reference level here is that of tetrapod fauna buried in the upper Inder Formation at the Kara-Bala-Kantemir site near the Lake Inder (Fig. 3; Ochev, 1973, 1987; Ochev and Smagin, 1974). The fauna includes mastodonsaurs of the Cis-Urals type and abundant "pustulate" *Plagioscutum caspiense* Shishkin that is morphologically more advanced than "vermiform" *P. ochevi* of the Donguz Formation (Shishkin, 1987). Hence, this fauna is an analog of the Bukobai *"Mastodonsaurus"* fauna and may be attributed, like the latter, to the upper Ladinian. The Inder tetrapods are obviously younger than the *Eryosuchus* fauna. According to this but contrary to the common opinion, at least the upper part of the Inder Formation (the *Pulviella aralsorica* Zone) should be younger than the Donguz Formation of the Cis-Urals.

Meanwhile, as mentioned above, the upper ostracod zone of the Inder Formation is dated back to the late Anisian–early Ladinian (Kukhtinov, 1999). Palynological data imply that the upper Kiil Formation, the age analog of above zone in the left bank of the Ural River, is either concurrent (Yaroshenko *et al.*, 2001) or the late Anisian (Il'ina, 2001). In the latest correlation scheme suggested for Triassic deposits in the Cis-Ural and Caspian region (Kukhtinov *et al.*, 2001), the entire Inder Formation is referred to the Anisian despite all the disagreements with respect to microfossil-based age interpretation of its upper part, which have not been taken into account (Table 2). Thus, estimating the Inder Formation stratigraphic range based on either vertebrate remains or microfossils, we get difference equal to one substage at least.

As an argument against correlation of the *Pulviella aralsorica* Zone (or its part) with the Ladinian Stage, we should mention a opinion that the tetrapod-bearing lens of greenish gray clays belongs in the Kara-Bala-Kantemir section not to the Inder Formation, but in fact to the lower part of the overlying Masteksai Formation (*Gemmanella schweyeri* Zone of the lower Fassanian). This opinion is based on conclusions of M.N. Shelekhova who argued that palynological assemblages from the bone-bearing lens and the Masteksai Formation are similar and correlative with the Bukobai assemblage of palynomorphs (Movshovich, 1998; Lozovsky *et al.*, 2002).

The Masteksai but not Inder age of the vertebrate fauna under consideration is doubtful however. First, Starozhilova found the ostracod assemblage typical of the *Pulviella aralsorica* Zone immediately above the bone-bearing lens of the Kara-Bala-Kantemir section (Ochev and Smagin, 1974). Second, immediately below the lens, dark clays with limestone interbeds, which undoubtedly belong to the Inder Formation, enclose rare remains of the same *"Mastodonsaurus"* fauna as in the lens. It is especially important that we found at this level not the "unidentifiable bones" (Movshovich, 1998, p. 23) but remains of "pustulate" plagiosaurine *Plagioscutum caspiense* (Shishkin, 1987, p. 9), which are stratigraphically significant (Shishkin and Ochev, 2002). All these data imply an insignificant age difference between the upper Inder and lower Masteksai strata and suggest diachronism of changes in ostracod and palynological assemblages in the considered interval of the section. The hiatus between the Inder and Masteksai formations and their analogs appears to be small (if there is any) or developed locally in opinion Kukhtinov.

Thus, stratigraphic ranges of the Middle Triassic units and position of the Anisian–Ladinian boundary are still problematic in the Caspian and the Cis-Urals regions (Fig. 3, Table 2). As for the former region, it is impossible at present to suggest a model that would reconcile the mentioned discordance in age estimates based on microfossils and vertebrate fauna.

Hence, the Ladinian age of the Bukobai Formation is adopted in both correlation models discussed above. Nevertheless, they suggest a limited but different stratigraphic range for the upper part of Inder Formation: the upper Anisian–lower Ladinian interval based on microfossils or to the upper Ladinian time span based on vertebrates.

A sharply different opinion was argued for in works by Movshovich (Movshovich and Kozur, 1975; Movshovich, 1980, 1998). This researcher did not extend the Inder Formation range higher than the middle Anisian and correlated the Bukobai Formation of the Cis-Urals (without its upper part) with the upper Anisian (Table 2). In other words, he suggested a much higher position of the Anisian–Ladinian boundary in the East European sections as compared with its levels in the other correlation schemes.

Conclusions of Movshovich are based on identifications of Triassic ostracods (mostly in Caspian sections), which have been done by Kozur, and on related generalized stages in development of the Middle Triassic

sedimentation in Europe. As he concluded, at the level of Inder–Masteksai boundary in the Caspian region (and correspondingly, of the Donguz–Bukobai boundary in the southern Cis-Urals) there is a hiatus spanning the upper Pelsonian–lower Illyrian and reflecting a regression episode recorded in the Middle Muschelkalk of the Germanic basin. Accordingly, the lower part of Masteksai Formation and the greater (or entire, as supposed earlier, Movshovich, 1980) interval of the Bukobai Formation have been attributed to the upper Anisian.

It is interesting that the same interval of Caspian sections (the Inder and Masteksai horizons), where Movshovich saw signs of the middle Illyrian regression, shows, in opinion of other stratigraphers, the records of progressing Illyrian–Fassanian transgression that culminated at the level of mid-Upper Muschelkalk of the Germanic basin (Lozovsky *et al.*, 2002). In this interpretation, the Inder and Masteksai formations are attributed to the upper Anisian and Fassanian, respectively.

It is obvious that conclusions of Movshovich are hardly compatible with views of other specialists based on distribution of different fossils, ostracods included, in the Middle Triassic sections of Eastern Europe. In the Caspian region, the Inder–Masteksai boundary cannot be older than the late Anisian according to the unanimous opinion (Kukhtinov, 1999; Ochev, 1999; Shishkin and Ochev, 1999; Il'ina, 2001; Yaroshenko *et al.*, 2001). In addition, neither ostracod nor vertebrate faunas suggest a noticeable hiatus between these two subdivisions (see above) and, hence, the record of regression (Lozovsky *et al.*, 2002).

Views of Movshovich are similarly incompatible with the known paleontological characterization of the Middle Triassic in the Cis-Urals, where Kozur identified ostracods from Borehole C-36 in the east of the Novocherkassk basin, 60 km eastward of the Donguz and Bukobai formations stratotype areas (Fig. 3). Basing on predominantly sediment coloration, Movshovich (1980), who followed geologists of the Gir'yal Geological Survey Team, attributed the gray colored deposits with ostracods to the Surakai Formation base above the Bukobai red beds. In opinion of Kozur, the ostracod assemblage he studied spans the interval from the upper Anisian (Illyrian) to the lower Ladinian (Fassanian).[7] The Bukobai Formation was considered therefore as corresponding to the upper Anisian (or upper Anisian–basal Fassanian in Movshovich, 1998, pp. 19, 23) and the Donguz Formation was consequently regarded as the older subdivision (Table 2).

Conclusions of Movshovich have been already commented (Shishkin and Ochev, 1992), but he insisted on them in his latest work (Movshovich, 1998) and we decided to present below some additional critical remarks. First, it is difficult to understand the reasons that compelled him to attribute the Bukobai Formation (or its larger part) to the Anisian and the Surakai Formation to the Ladinian. It is general opinion that the Bukobai Formation stratotype yields only the Ladinian assemblages of palynomorphs (Makarova, 1975; Tuzhikova, 1975, 1979; Makarova and Vergaĭ, 1995; Yaroshenko *et al.*, 2001), concurrent vertebrate remains (Shishkin and Ochev, 1992), and the Ladinian–Carnian flora (Dobruskina, 1980), whereas palynomorphs of the Surakai Formation are not older than the Carnian (Tuzhikova, 1975). These data logically imply that the ostracod assemblage from the Novocherkassk basin should be older than the Surakai Formation, and that the gray ("Surakai") color of rocks, which is characteristic of different Middle Triassic levels in sections of the Cis-Urals, cannot be used in itself as an age indicator.

For instance, Tverdokhlebov and Makarova (1976) who studied stratigraphy of disjunctive basins in the southern Cis-Urals clearly showed that the lower limit of predominantly gray deposits characteristic of the Surakai Formation is diachronous, slipping down and crossing the Bukobai and even Donguz horizons. Critical remarks of Tuzhikova (1975) cited by Movshovich (1980) only doubted their observation without concrete disproving arguments. The arguments in favor of that observation have been obtained by drilling in a peripheral area of the Tamarkuduk basin, the Sol'-Ilets district of the Orenburg oblast (Borehole 11-c Fig. 3). I.S. Makarova who studied core samples collected by V.A. Efremov showed that gray-colored deposits, which were previously referred to the Middle Jurassic (depth level of 60 m), yield the Bukobai palynological assemblage. In deposits of the same color (formerly Upper Triassic) from the depth of 142 m, she identified palynomorphs characteristic of the Donguz Horizon.

Finally, Movshovich attributes the Bukobai Formation to the Anisian Stage referring to presence of Illyrian–Fassanian ostracods at the "Surakai" Formation base in the Novocherkassk basin. However, this conclusion is in obvious disagreement with the fact that the Bukobai (Ladinian) spore–pollen assemblage found in the basin is similar in composition to palynomorphs typical of the formation stratotype. The assemblage studied by Makarova includes characteristic Bukobai species *Minutosaccus potoniei* Madl., *M. acutus* Madl., *Carnisporites mesozoicus* Klaus, *Acanthotriletes ilekensis* Kopyt., *Florinites walchius* Kopyt., *F. pseudostriatus* Kopyt., *Alisporites australis* Jersey, *Sulcatisporites reticulatus* Scheuring, *S. sulcus* Klaus and was found at the depth level of 55 m in Borehole 17 drilled by the Gir'yal Geological Survey Team (Fig. 3). Palynological assemblage typical of the Donguz Formation was detected here at the depth of 87 m.

It is difficult to understand also why Movshovich (1998) thinks that work by Dobruskina (1980) on Middle–Upper Triassic *Scytophyllum* flora from the Cis-Urals supports the Ladinian age of the Surakai Forma-

[7] Kukhtinov suggests the Fassanian interval for the assemblage.

tion. That flora is of a wider Ladinian–Carnian range and cannot help to figure out a narrower time span for sediments in question. In reality, the Late Triassic age of the Surakai Formation is substantiated by palynoflora found in this subdivision.

In addition, all the units of the Caspian region (the lower Sarpinskii Subformation and the upper part of the Masteksai Formation in the Krasnoyar dome area; Starozhilova and Shelekhova, 1987), which have been placed by Movshovich at the level the "Surakai" ostracod assemblage from the Novocherkassk basin of the Cis-Urals are correlative, in opinion of many micropaleontologists (Kukhtinov, Yaroshenko, and others), with the Bukobai Horizon of the Cis-Urals. Exactly this variant of correlation has been adopted in 1999 at the meeting on Triassic stratigraphy of the Caspian syneclise.

In general, it is clear that the key problem of the Middle Triassic stratigraphy in Eastern Europe lies in surmounting discordances between age determinations based on microfossils and vertebrates remains. It would be purposeless to extend the "interval of uncertainty" taking into account the conclusions of Movshovich (1998), which contradict all the other data.

Differences in stratigraphic position of boundaries between the Middle Triassic horizons of the southern Cis-Urals and between the Inder and Masteksai horizons of the Caspian region are undoubtedly related to subjective interpretation of some facts considered above. Each of the fossil groups studied has merits and disadvantages. Frequently occurring ostracods are facies-dependent. The long-distance correlations may be incorrect because of asynchronous appearance of particular ostracod taxa in different regions, may have concurrent ranges in some region and succeed each other in the other sections (Kukhtinov, 1999). Vertebrate remains open a wide perspective for using the evolutionary method, but they are often fragmentary and occur within very limited stratigraphic intervals. Moreover, the method of comparative evolutionary levels does not exclude a possibility of heterochronous parallelism, as the same forms could have diverging rates of evolution in different biogeographic areas. Palynomorphs occur in a more complete successions, but changes in their assemblages are less distinct than in tetrapod faunas.

Difficulties in solving the problem under consideration may be also connected with miscorrelation between Triassic deposits of the Germanic basin and Alpine region. Correlating the East European Triassic deposits with either the Alpine, or the Germanic basin scales, one will get different results in such a case.

A real way to solve this problem is to study in detail successive changes in all leading fossil groups useful for dating the East European Middle Triassic. This approach is promising much with respect to palynostratigraphy of the Bukobai Formation in the southern Cis-Urals. To rapidly establish the complete successions of diagnostic vertebrate remains is a more difficult task in the Cis-Urals and West Europe.

At the present stage of knowledge, we have no reliable base to correlate unambiguously the Middle Triassic subdivisions of the Caspian region and southern Cis-Urals with the Anisian and Ladinian stages, as it was attempted at the meeting of 1999.

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Palynological Assemblages of Continental Lower Triassic in Eastern Europe and Their Interregional Correlation. Paper 1: Palynological Assemblages of the Induan Stage

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Abstract—Four palynological assemblages characteristic of the Nedubrovo, basal Astashikha, middle–upper Astashikha–Ryabi, and Sarafanikha members of the Vokhma Formation are distinguished in Lower Triassic continental deposits of the Moscow syneclise. Palynological assemblages of similar composition are known from continental deposits in other areas of the East European platform and from marine sediments bearing ammonites and conodonts. Palynological assemblages *Cycadopites* sp.–*Klausipollenites schaubergeri* and *Striatoabieites richteri–K. schaubergeri* are correlated with the *Otoceras concavum* Zone of the ammonite scale. The *Densoisporites complicatus–Ephedripites* sp. assemblage spans the range of the *Otoceras boreale* Zone, and the *Ephedripites permasensis–Pechorosporites disertus* assemblage is typical of the *Ophiceras commune* Zone and probably of a higher part of the Induan Stage. The established correlation is consistent with data obtained by ecotone and paleomagnetic methods.

Key words: Lower Triassic, Induan Stage, palynological assemblages, correlation, Moscow syneclise.

INTRODUCTION

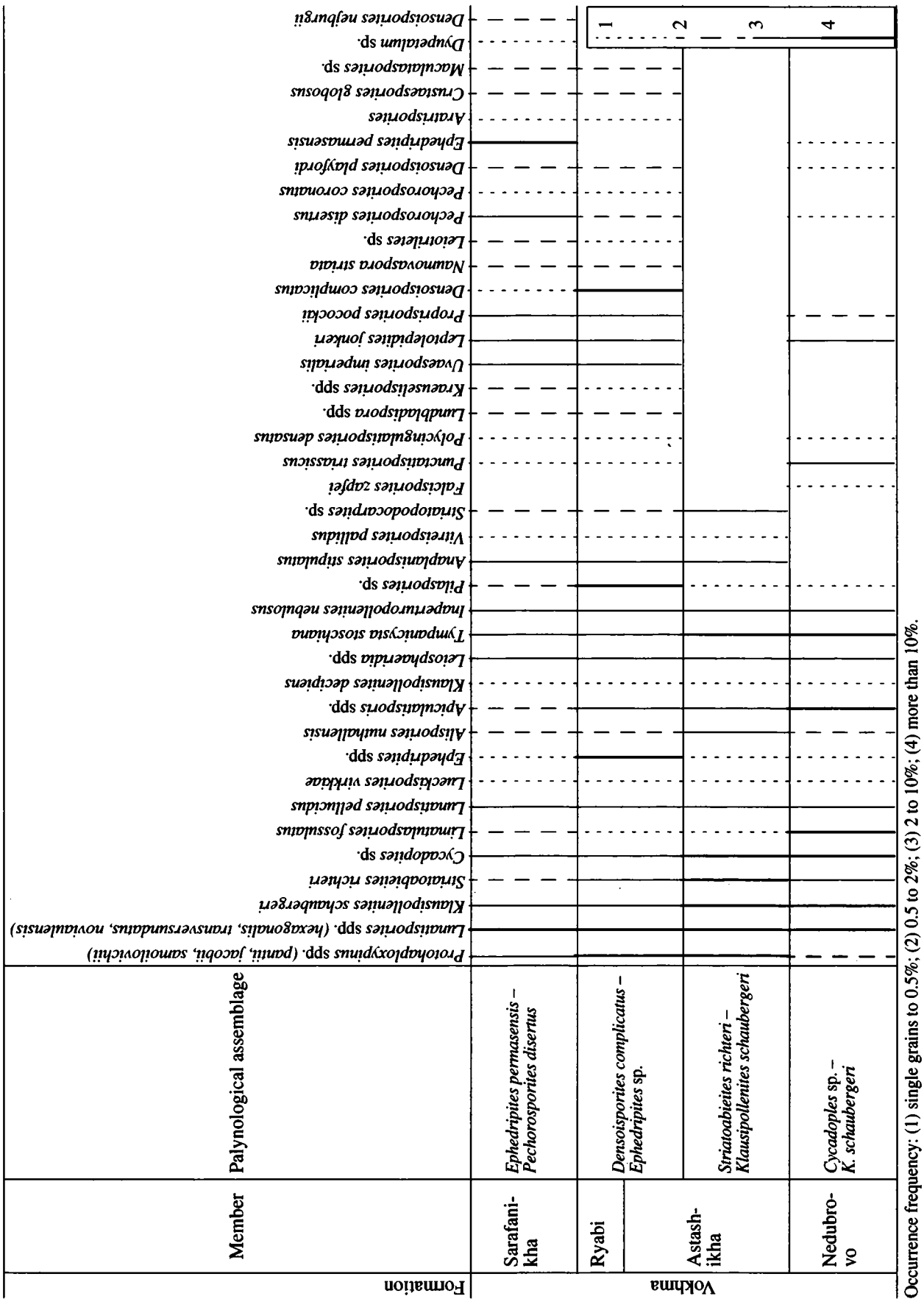
In continental deposits of eastern Europe, the most complete palynomorph assemblages correlative with palynozones of marine deposits of the Induan Stage are known from the basal interval of the Vetluga Group (Superhorizon) of the Moscow syneclise. Kyuntsel' (1965, 1969) and Goman'kov *et al.* (1986) were first to distinguish this microflora in gray mudstones and siltstones of the Ryabi Horizon (a member at present) of the Vokhma Formation, the lower one in the Vetluga Group. The "first and oldest assemblage of spores and pollen," which was found here, turned out to be comparable with that of the Toad/Grayling Formation of western Canada (Jansonius, 1962).

Later on, Yaroshenko reexamined preparations of Kyuntsel' and published together with Lozovsky, who collected new materials in the Moscow syneclise, the additional palynological characterization of the Vokhma Formation basal part (Lozovsky and Yaroshenko, 1994; Yaroshenko and Lozovsky, 1997). Three individual assemblages correlative with microfloras known from the Induan Stage lower half were named subsequently after their characteristic palynomorph taxa. These were the *Striatoabieites richteri–Klausipollenites schaubergeri* assemblage from the base of the Astashikha Member, the *Densoisporites complicatus–Ephedripites* sp. assemblage from the middle–upper Astashikha and Ryabi members (the lower Vokhma Subformation), and the *Ephedripites permasensis–Pechorosporites disertus* assemblage from the Sarafanikha Members of the upper (Krasnye Baki) subformation of the Vokhma Formation (Yaroshenko and Goman'kov, 1998). A recent series of works (Krassilov *et al.*, 1999a, 2000; Afonin, 2000; Lozovsky *et al.*, 2001; Lozovsky *et al.*, 2001b) was aimed to characterize palynomorphs from the basal Nedubrovo Member of the Vokhma Formation (Lozovsky, 2001; Lozovsky *et al.*, 2001b), which were attributed by Afonin based on dominants to the *Cycadopites* sp.–*Klausipollenites schaubergeri* palynological assemblage. Palynomorph assemblages comparable in composition with some of the aforementioned assemblages from the Lower Triassic of the Moscow syneclise are also know from the North Caspian depression (Makarova and Romanovskaya, 1983, 1999), Timan–Pechora region (Varyukhina, 1971, 1986; Varyukhina and Golubeva, 1983; Tuzhikova, 1985; Yaroshenko *et al.*, 1991), and on the Novaya Zemlya Islands (Fefilova, 1999).

CHARACTERIZATION OF PALYNOLOGICAL ASSEMBLAGES

The *Cycadopites* sp.–*Klausipollenites schaubergeri* palynological assemblage is the oldest one in the Lower Triassic of Eastern Europe (Table 1). It is typical of the Nedubrovo Member exposed near the eponymous village on the left side of the Kichmenga River (left tributary of the Yug River). The assemblage was recovered from a bed of old-oxbow ribbon to laminated alluvium

Table 1. Stratigraphic ranges of main palynomorph taxa in successive members of the Vokhma Formation, Moscow syneclise



(Afonin, 2000) composed of gray to greenish gray flaggy mudstone and siltstone. As is established, there are eolian ash particles in this bed (Lozovsky *et al.*, 200la, 2001b).

Characteristic of the assemblage is presence of the Upper Permian bisaccate pollen that includes both the non-striate and striate forms, the former represented by *Klausipollenites schaubergeri* (Potonié et Klaus) Jansonius, *Alisporites nuthallensis* Clarke, and *Falcisporites zapfei* Potonié et Klaus, and the latter by *Lueckisporites virkkiae* Potonié et Klaus, *Striatoabieites richteri* (Klaus) Hart, and *Lunatisporites noviaulensis* (Leschik) Foster. The listed taxa coexist with Triassic species *L. pellucidus* (Goubin) Helby, *L. transversundatus* (Jansonius) Fisher, and with associated spore taxa *Apiculatisporis* sp., *Limatulasporites fossulatus* (Balme) Helby et Foster, *Punctatisporites triassicus* Schulz, *Polycingulatisporites densatus* (de Jersey) Playford et Dettmann, *Leptolepidites jonkeri* (Jansonius) Yaroshenko et Golubeva, *Proprisporites pocockii* Jansonius, and *Densoisporites playfordi* (Balme) Dettmann. The assemblage is named because of a prime significance of *Cycadopites* and *C. schaubergeri* representing more than 50% of all palynomorphs.

Next in abundance is pollen of *L. pellucidus* and *S. richteri.* Species *Protohaploxypinus* cf. *pantii* (Jansonius) Orlowska-Zwolinska, *L. virkkiae*, and *L. transversundatus* are less abundant, whereas *Pechorosporites disertus* Yaroshenko et Golubeva, *Ephedripites permasensis* Yaroshenko, *Striomonosaccites* sp. and *Triadispora* sp. occur as rare specimens. Miospores listed above occur in association with remains of planktonic organisms *Pterospermella, Pilasporites, Tympanicysta stoschiana* Balme, and *Inaperturopollenites nebulosus* Balme. The assemblage was described in detail in several works (Krassilov *et al.*, 1999a; Afonin, 2000; Lozovsky *et al.*, 2001a, 2001c). Afonin who described morphological anomalies of saccate pollen taxa *Protohaploxypinus ?, Lunatisporites ?* and *Klausipollenites* defined them as a distorted orientation of striae (taeniae) and as development of additional air sacs responsible for appearance of three-, four- and even five-lobed pollen grains. These anomalies likely developed under influence of an intense ultraviolet irradiation because of a hole in ozone layer, which originated in response to volcanic eruptions at the deposition time of the Nedubrovo Member.

The *Striatoabieites richteri–Klausipollenites schaubergeri* palynological assemblage characterizes a lower part of the Astashikha Member (Yaroshenko and Lozovsky, 1997; Yaroshenko and Goman'kov, 1998). It was established in one core sample of Borehole 115 (depth level of 129.6 m) near the village of D'yakonovka, the Soligalich area of the Kostroma oblast. Dominant in the assemblage is striate pollen of *S. richteri, Striatopodocarpites* spp., and less abundant *P. pantii, P. jacobii* (Jansonius) Hart, *P. samoilovichii* (Jansonius) Hart, *L. pellucidus, L. noviaulensis*, and *L. transversundatus.* In the group of non-striate pollen taxa, dominant *K. schaubergeri* is associated with subordinate *Klausipollenites* sp., *Alisporites* spp., *Platysaccus* sp. and *Vitreisporites* sp. The monocolpate pollen is represented by that of *Cycadopites* (more than 10%) and rare *Monosulcate* sp. A. Mangerud. Spores are of a low abundance and not diverse, being represented by small trilete forms *Apiculatisporis* sp., *A. lanjouwii* Jansonius, and *Anaplanisporites stipulatus* Jansonius (about 10% in sum of the total number of palynomorphs). Specimens of *Ephedripites* sp., *Vittatina* sp., *Praecolpatites* sp., and *Aumancisporites* sp. are rare. Associated planktonic organisms are *T. stoschiana, Leiosphaeridia, Pilasporites*, and *I. nebulosus.* The eponymous taxa of the assemblage are the abundant *S. richteri* and *K. schaubergeri* representing over 40% of all palynomorphs.

In general, the assemblage under consideration is comparable with the *Cycadopites* sp.–*K. schaubergeri* assemblage described higher. Dominant in both are *Klausipollenites* forms, mostly *K. schaubergeri*, and *Cycadopites* sp., which associate with representatives of genera *Protohaploxypinus, Striatoabieites, Lunatisporites, Lueckisporites, Platysaccus* and *Alisporites*, and with spores of the *Apiculatisporis* type. At the same time, spores of the older assemblage are more diverse, but the material used for discrimination of the *S. richteri–K. schaubergeri* assemblage was limited and this could be an explanation for the mentioned distinction. Nevertheless, miospores of this assemblage were regarded as representative, clearly different in composition from microfloras of overlying members, and had been attributed to the oldest assemblage of the Vokhma Formation before discrimination of the *Cycadopites* sp.–*Klausipollenites schaubergeri* assemblage characterizing the formation basal part. A great similarity between two assemblages brought into challenge stratigraphic validity of the *S. richteri–K. schaubergeri* assemblage, which was distinguished in a single sample and can be reinstated, despite the priority principle, only after examination of additional materials. Characteristic of both assemblages are also *Leiosphaeridia, Pilasporites, I. nebulosus*, and *T. stoschiana.* The last taxon is now classed with a green alga (Krassilov *et al.*, 1999b; Afonin *et al.*, 2001) in contrast to the former opinion of paleontologists that it represents fungal remains.

The *Densoisporites complicatus–Ephedripites* sp. palynological assemblage is known from several sections of Lower Triassic deposits in the Moscow syneclise. The assemblage stratigraphic range corresponds to the middle and upper parts of the Astashikha Member, which have been penetrated by drilling near the villages of Okatovo (Borehole 185, depth level of 132 m) and Anyug (Borehole 5, depth level of 160) in the Kologriv district, and near the village of Ryabinovtsy (Borehole 86, depth interval of 90–94 m) in the Pyshchug area, the Kostroma oblast. In addition, the same assemblage is

characteristic of the Ryabi Member, Exposure 42 near the village of Sholga on the right side of the Yug River, the Podosinovskii district of the Vyatka oblast. The assemblage is more diverse than assemblages described above. It includes trilete cavate spores *Densoisporites complicates* Balme and *D.* ex gr. *complicates* (20–30%) associated with polypericolpate pollen *Ephedripites* spp., *E. steevesii* (Jans.) de Jersey et Hamilton and other species of the genus included, which represent up to 30% of the assemblage in some samples. In addition to abundant *D. complicates* and *D.* ex gr. *complicates*, there are first appearing spores *Kraeuselisporites saeptatus* Balme, *Kraeuselisporites* sp., *Lundbladispora willmotti* Balme, *L. brevicula* Balme, *L. obsoleta* Balme, *Uvaesporites imperialis* (Jansonius) Utting, *Naumovaspora striata* Jansonius, *Pechorosporites coronatus* Yaroshenko et Golubeva, *Camptotriletes* sp., *Rewanispora* sp., *Aratrisporites* sp., and *Crustaesporites globosus* Leschik., all missing from older palynological assemblages. The other taxa first appearing in the Nedubrovo Member are *D. playfordi, P. pocockii, L. jonkeri, P. disertus*, and *L. fossulatus.* Components inherited from older assemblages are spores *Apiculatisporis* spp. and *Anaplanisporites stipulates* and pollen taxa *P. pantii, P. jacobii*, and *Lunatisporites* spp., taeniate forms of the latter being slightly more abundant than before. *S. richteri* and *K. schaubergeri*, as well as single specimens of *L. virkkiae, Vittatina, P. triassicus, Osmundacidites* sp., *Endosporites papillatus* Jansonius, *Chordasporites* sp., and *Pretricolpipollenites* sp., still occur at this level. Inaperturate palynomorphs, such as *I. nebulosus, Maculatasporites, T. stoschiana, Leiosphaeridia* sp. and *Pilasporites* sp. are abundant, especially the last two taxa. Being of specific composition, the *D. complicatus–Ephedripites* sp. assemblage has nevertheless certain features characteristic of many Lower Triassic palynofloras known in other regions of the world.

The *Ephedripites permasensis–Pechorosporites disertus* palynological assemblage is typical of the Sarafanikha Member penetrated by Borehole 16 (depth interval of 90.6–91.7 m) near the village of Permas, the Nikol'skii district of the Vologda oblast. In distinction from the previous assemblage, it is enriched in pollen *Ephedripites permasensis* (18%) whose single grains were found in the lower *Cycadopites* sp.–*K. schaubergeri* assemblage. Remarkable characters of the assemblage consist in the first occurrence of *Densoisporites nejburgii* (Schulz) Balme and *Dyupetalum* sp. and in a higher abundance of *P. disertus* and *L. fossulatus.* In addition, the abundance rate of *D. playfordi, Densoisporites* sp., *Lundbladispora* spp., *Kraeuselisporites* spp., *U. imperialis, A. stipulatus, L. jonkeri, P. pocockii*, and *N. striata* is also greater. Among the bisaccate pollen, there are many species of the previous assemblages, which are represented by striate, especially by more abundant taeniate forms, and by non-striate taxa. *Klausipollenites* spp. and *K. Schaubergeri*, both well preserved, are still present among the latter, whereas abundance of striate *S. richteri* is noticeably lower, although *Cycadopites* forms remain to be important components of the assemblage. Quantity of planktonic *Leiosphaeridia* sp., *Pilasporites* sp., *I. nebulosus, T. stoschiana*, and *Maculatasporites* is somewhat decreased. Spores *D. complicatus, D.* ex gr. *complicates* and diverse *Ephedripites* pollen, the components dominant in the previous assemblage, are rare, and this is a distinctive feature of the *E. permasensis–P. disertus* assemblage. It is also interesting that first specimens of *Pleuromeia* sp. are recorded at the same stratigraphic level in the Yug River basin (Lozovsky *et al.*, 2001b).

Four palynological assemblages described above characterize the greater part of the Vokhma Formation in the Moscow syneclise exclusive the upper Anisimovo Member of this subdivision. They represent in sum the most complete succession of the Lower Triassic palynomorphs not only in Eastern Europe but, probably, in the entire Laurasia.

CORRELATION WITH PALYNOZONES AND PALYNOLOGICAL ASSEMBLAGES OF CONTINENTAL DEPOSITS

Two lower palynological assemblages are very similar and thus suitable only for a joint correlation in continental sections. They both reveal a certain similarity with miospore fauna from sediments buried between two basaltic flows in the Chernyshev Chain of Hills (Tuzhikova, 1985), where these sediments were penetrated by Borehole SDK 458 (depth interval of 181.1–168.8 m). The similarity consists in a high percentage of bisaccate striate and non-striate pollen, especially of *K. schaubergeri*, in presence of several taxa in common, and in characteristic morphological anomalies of saccate pollen grains, among which there are grains having three air sacs and forms with sacs closely arranged near the proximal pole (Tuzhikova, 1985, p. 66).

Both assemblages from the Moscow syneclise are also comparable with the Subassemblage "A1" from the lower Admiralteiskaya Subformation of the Novaya Zemlya Islands (Fefilova, 1999). The subassemblage consisting of mixed Upper Permian and Lower Triassic palynomorphs was correlated by Fefilova with the early Griesbachian assemblage "P" of the Barents Sea shelf (Hochuli *et al.*, 1989).

The next *D. complicatus–Ephedripites* sp. assemblage of the Moscow syneclise is similar in composition to miospores from sedimentary member underlying basalts in the Korotaikha depression of the northern Cis-Urals[1] (Tuzhikova, 1985). Assemblages of the

[1] In the Timan–Pechora region, there are likely three basaltic horizons, the lower one of the Late Permian and two upper of the Early Triassic (Lozovsky and Balabanov, 1998). Deposits occurring below basalts in the Korotaikha depression are younger than sediments sandwiched between basalts in the Chernyshev Chain of Hills.

Moscow syneclise and Cis-Urals are comparable because of their very diverse taxonomic composition. Despite some uncertainty in taxa identification, spores of the Cis-Uralian assemblage are more diverse however and include characteristic cavate forms *D. playfordi, D. complicatus, K. saeptatus, L. brevicula* and *L. willmotti*. Pollen groups of both assemblages are also comparable, including nearly all the species of the genus *Ephedripites* (*Gnetaceaepollenites*), which were originally described from the Toad/Grayling Formation of western Canada (Jansonius, 1962), and abundant pollen grains of striate and non-striate morphology. The similarity is emphasized by a high content of some algal taxa, e.g., of *Pilasporites*.

In distinction from the assemblage of the northern Cis-Urals, concurrent miospores from the Ershov Horizon of the North Caspian syneclise are not so diverse (Makarova and Romanovskaya, 1999). Nevertheless, the horizon yields *Calamospora* sp., *Densoisporites* sp., *Punctatisporites triassicus, Limatulasporites fossulatus, Cycadopites* sp., *Lunatisporites noviaulensis, Protohaploxypinus samoilovichii, P. jacobii, P. chaloneri* Clarke, *I., virkkiae, K. schaubergeri*, abundant *Ephedripites* spp., and *T. stoschiana* whose assemblage is presumably correlative with the *D. complicatus–Ephedripites* sp. assemblage of the Moscow syneclise.

The *Densoisporites complicatus–Ephedripites* sp. assemblage of the Moscow syneclise can be correlated with the *Lundbladispora obsoleta–Protohaploxypinus pantii* Palynozone in the Baltic Group of the Lower Buntsandstein in Poland (Orlowska-Zwolinska, 1984; Fijalkowska, 1994, 1995; Fijalkowska-Mader, 1998, 1999). In addition to features in common, the assemblage of miospores from that palynozone simultaneously shows some distinctions. Both assemblages include abundant cavate spores of lycopsids, such as *Lundbladispora* spp., *Densoisporites* spp., *Kraeuselisporites* spp., and striate pollen *S. richteri, P. pantii*, and *L. noviaulensis* in association with abundant *Ephedripites (Gnetaceaepollenites), Cycadopites* sp., and *T. stoschiana*. Nevertheless, the palynological assemblages under comparison differ from each other in relative abundance of these and some other taxa. *Lundbladispora* spp., especially *L. obsoleta, L. brevicula* and *L. willmotti*, are more abundant in Poland than in the Moscow syneclise, where species *D. complicatus* unknown so far in sections of Poland is dominant. Moreover, the assemblage from Poland includes acanthomorphic acritarchs and shows a somewhat greater abundance of *E. papillatus, Cyclotriletes*, and *Punctatisporites*. Megaspores of the *Otynisporites eotriassicus* Zone, which occur near the Baltic Group base, were also found in the Nedubrovo member (Krassilov *et al.*, 1999a; Lozovsky *et al.*, 2001), and this implies that basal levels of the Baltic and Vetluga groups may be concurrent. It is possible to suggest also that the assemblage of the *L. obsoleta–P. pantii* Palynozone is of a wide stratigraphic range enclosing the ranges of older *Cycadopites* sp.–*K. schaubergeri* and *S. richteri–K. schaubergeri* assemblages of the Moscow syneclise, as one can judge from the close composition of pollen despite a greater abundance of spores.

Only *D. playfordi, Lunatisporites*, and *Cycadopites* forms suggest an approximate similarity between the *D. complicatus–Ephedripites* sp. assemblage from the Moscow syneclise and the *Endosporites papillatus–Densoisporites playfordi* assemblage from stratigraphically concurrent, though more remote deposits of the lower Buntsandstein of Germany. The main distinction is a much higher abundance of acritarchs in the latter (Reitz, 1985; Schulz, 1998, written communication). Accordingly, the Buntsandstein of Germany was deposited in a deeper setting.

The fourth *Ephedripites permasensis–Pechorosporites disertus* palynological assemblage of the Moscow syneclise is correlative with the *P. disertus* assemblage from the lower part of the Charkabog Formation of the Pechora syneclise (Yaroshenko *et al.*, 1991). The former differs from the latter mainly in abundance and diversity of *Aratrisporites* and *Pechorosporites* forms: in the Moscow assemblage spores *Aratrisporites* sp. are rare and *P. disertus* is of a low abundance (1 to 5%), while in the Pechora assemblage they are abundant, being represented by several species.

According to percentages of *Pechorosporites* spp., *Kraeuselisporites* spp., *Densoisporites* spp., *Lundbladispora* spp., *P. pocockii, Polycingulatisporites, L. pellucidus, L. transversundatus, Klausipollenites* spp., *Cycadopites* sp., *Pilasporites* and some other taxa, the *E. permasensis–P. disertus* assemblage of the Moscow region is comparable with subassemblage "A_2" (and partially with "A_3") from topmost horizons of the Admiralteiskaya Formation of the Novaya Zemlya Islands (Fefilova, 1999), where they are enriched in *K. trisporites* spores. The last peculiarity was regarded as important, suggesting that two assemblages are correlative with the *Pechorosporites disertus* assemblage of the Pechora syneclise (Yaroshenko *et al.*, 1991).

CORRELATION WITH PALYNOZONES IN MARINE DEPOSITS

As is known, correlation between marine and continental formations is based on different methods: the indirect stratigraphic, palynological, paleomagnetic, paleogeographic, and ecotone ones. The palynological method is of prime importance therewith and Menner (1962) regarded it as most suitable for the direct correlation of contrasting sedimentary facies. Exactly this method, used for correlation of continental deposits in Eastern Europe with marine standards, enabled recognition of the Induan Stage in the former.

Palynological assemblages of the Moscow syneclise can be convincingly correlated with their counterparts from the Induan marine sediments of the Boreal basin (Greenland, Spitsbergen, Arctic Canada, Barents Sea)

that was directly connected with the Moscow basin (Lozovsky, 1994).

The connections between two basins are detectable based on distribution of both the miospores in common and the planktonic organisms *Leiosphaeridia, T. stoschiana*, and *Pilasporites*, which dwelt in different settings from marshland–lacustrine and fluvial to deep-water marine ones. The continental and marine palynological assemblages differ from each other because of a very low abundance or complete absence of acanthomorphic acritarchs in the former (Jansonius, 1962; Balme, 1979; Hochuli *et al.*, 1989; Mangerud, 1994; Utting, 1994; Vigran *et al.*, 1998; Mørk *et al.*, 1999).

The oldest palynological assemblages known from Triassic sections are transitional in composition. Typical of them is coexistence of the Late Permian (Zechstein) taxa *K. schaubergeri, A. nuthallensis, S. richteri, Lueckisporites virkkiae, L. noviaulensis, Vittatina*, and the Early Triassic *Protohaploxypinus samoilovichii, P. pantii, P. jacobii, L. pellucidus, L. transversundatus*, and *L. noviaulensis*. The lower *Cycadopites* sp.–*K. schaubergeri* and *S. richteri–K. schaubergeri* palynological assemblages (from the Nedubrovo and basal Astashikha members of the Moscow syneclise, respectively) appear to be of the sort. They show some similarity with the oldest Griesbachian assemblages from the *Otoceras concavum* Zone of the ammonite scale (Table 2). For instance, the content of *Protohaploxypinus* pollen, taeniate and other striate forms, e.g., of *S. richteri*, and relatively impoverished composition of spores suggest that palynological assemblages of the Moscow syneclise are quite close in composition to the *Protohaploxypinus* assemblage of eastern Greenland (Balme, 1979) despite a higher content of pollen *Ephedripites* spp. rarely occurring in the lower horizons of the Moscow sections. In Greenland, host deposits of the assemblage bear ammonites *Otoceras* sp. and *Glyptophiceras (Hypophiceras) triviale* Spath. In opinion of many researchers, the associated Permian corals, bryozoans and brachiopods are redeposited. Tozer, (1967) correlated the local *triviale* Zone with the basal interval of the *concavum* Zone that is the lowermost in the Griesbachian Stage of Canada and corresponds to the lower Induan Substage of the general Triassic scale.

Based on a rather high abundance of striate pollen and on associated spores, alga *T. stoschiana* and relic forms of the Upper Permian, which are characteristic of both lower assemblages of the Moscow syneclise, they can be correlated with the lower part of the *L. obsoleta–T. stoschiana* Palynozone in the lower Griesbachian Deltadalen Member of the Vardebukta Formation of Spitsbergen and Barents Sea (Hochuli *et al.*, 1989; Mangerud, 1994; Mørk *et al.*, 1999). Basal beds of the Deltadalen Member bear an impoverished assemblage of Permian, mostly redeposited pollen, which is analogous to pollen occurring in underlying deposits of the Kapp Starostin Formation. It can be suggested for sure that the *L. obsoleta–T. stoschiana* assemblage of Spitsbergen and Barents Sea is of a wider stratigraphic range, as it is shown below.

Several species in common (*S. richteri, P. samoilovichii, L. noviaulensis, T. stoschiana*) imply that lower palynological assemblages of the Moscow region are comparable with the lower Griesbachian assemblage of the *Tympanicysta stoschiana–Striatoabieites richteri* Palynozone distinguished at the base of the Blind Fiord Formation in the Sverdrup basin of Arctic Canada (Utting, 1994).

Comparing the older palynological assemblages of the Moscow syneclise with Induan palynoflora from the Toad/Grayling Formation of western Canada (Jansonius, 1962), one can easily note a great similarity in composition of taxa representing the diverse striate and taeniate bisaccate pollen. At the same time, the Canadian assemblage enriched in *Ephedripites* (*Gnetaceaepollenites*) pollen is comparable as well with the younger *D. complicatus–Ephedripites* sp. assemblage of the Moscow syneclise, in which the *Ephedripites* pollen is associated with principal spore taxa described in western Canada (Jansonius, 1962), where *T. stoschiana*, another species in common for two assemblages, was also found (Mangerud, 1994, p. 334), being formerly missed by Jansonius. The Canadian assemblage that was first distinguished in the Lower Triassic marine deposits seems to be of the composite type to a certain extent and spans most likely a considerable interval of the Induan Stage.

The third *D. complicatus–Ephedripites* sp. assemblage of the Moscow syneclise is closer in composition to the second *Lunatisporites* (*Taeniaesporites*) assemblage of eastern Greenland (Balme, 1979), than to the older *Protohaploxypinus* palynoflora. Rather abundant *L. noviaulensis, L. pellucidus, L. transversundatus, Ephedripites*, and *Cycadopites*, which are present in two indicated assemblages, associate with representatives of genera *Densoisporites, Lundbladispora*, and *Kraeuselisporites* and with persistently occurring *T. stoschiana.* However, in contrast to the Canadian assemblage, trilete cavate spores are more abundant and diverse in the Moscow assemblage. Deposits of eastern Greenland, which yield the *Lunatisporites* assemblage, bear the early Griesbachian ammonites *Otoceras* sp. and *Hypophiceras* sp. in association with bivalves *Claraia* sp. (Balme, 1979). The last group of fossils suggests a higher stratigraphic level of palynoflora.

With all its peculiarities, the *D. complicatus–Ephedripites* sp. assemblage of the Moscow region, which includes abundant spores and diverse *Ephedripites* pollen, is correlative with the upper part of the *L. obsoleta–T. stoschiana* Palynozone in the lower Griesbachian deposits in the Barents Sea southwestern shelf (Mangerud, 1994). That palynozone, like the Moscow assemblage, reveals coexistence of diverse pollen (*Protohaploxypinus* spp., *Lunatisporites* spp., *S. richteri, L. virkkiae, Klausipollenites* and various *Ephedripites*

species) with abundant and diverse spores. Among the latter, there were identified *D. playfordi, Lundbladispora, Kraeuselisporites* (*K. saeptatus* included), *U. imperialis, N. striata, P. pocockii, A. stipulatus,* and *Maculatasporites* forms. An important associated taxon is *T. stoschiana.* It is also important for correlation that rather abundant species *U. imperialis* were found in the Sholga Village section, where they frequently occur in quaternions like in the *L. obsoleta–T. stoschiana* Palynozone also rich in this taxon (Mangerud, 1994).

Palynomorphs of *L. obsoleta–T. stoschiana* Palynozone are comparable with the lower Griesbachian *Lunatisporites* (*Taeniaesporites*) assemblage of eastern Greenland and have been correlated with the assemblage "P" of the Barents Sea region (Hochuli *et al.*, 1989; Mørk *et al.*, 1999). Ammonites *Otoceras boreale* Spath present in upper strata of the unit that yields the last assemblage and bears as well *Tompophiceras* cf. *gracile* (Spath), bivalves *Claraia stachei* Bittn., *Anodontophora* sp., and gastropods *Bellerophon borealis* Spath and *Worthenia* sp. undoubtedly mean that host deposits of these fossils belong to the *boreale* Zone of the ammonite scale (Mørk *et al.*, 1999). Indications, which point to similarity between the *D. complicatus–Ephedripites* sp. assemblage and palynoflora of the Induan Toad/Grayling Formation of western Canada (Jansonius, 1962), are the diversity of *Ephedripites* pollen and a series of spore species in common. The latter were first described from the Canadian assemblage, although the trilete cavate spores are not as abundant here as in the Moscow region.

The *E. permasensis–P. disertus* assemblage characterizing the uppermost part of the Induan Stage in the Moscow syneclise is correlative with the *Densoisporites nejburgii–Proprisporites pocockii–Maculatasporites* sp. Palynozone in the Svalis-1 Formation of the central Barents Sea, in which palynomorphs occur in association with late Griesbachian ammonites of the *commune* Zone (Vigran *et al.*, 1998; Mørk *et al.*, 1999). All index species known from the Barents Sea region and associated *Aratrisporites, P. disertus, Lundbladispora* spp., *Kraeuselisporites* spp., *L. fossulatus, Lunatisporites* sp., *L. noviaulensis, L. pellucidus, Protohaploxypinus* spp. and *Cycadopites* sp. are present as well, though in different proportions, in the assemblage from Moscow sections. In the Barents Sea deposits, for instance, *Aratrisporites* spp. and *D. nejburgii* are dominant taxa and finds of *P. disertus* are rare. In distinction, the Moscow assemblage includes single specimens of the first taxon and a low quantity of two others. Vigran *et al.* (1998) argued that palynoflora from the Svalis-1 Formation is comparable with the *P. disertus* assemblage from the lower part of the Charkabog Formation of the Pechora syneclise (Yaroshenko *et al.*, 1991). The latter assemblage is close in composition, as noted above, to the *E. permasensis–P. disertus* assemblage of the Moscow region. The low occurrence frequency of index species *P. disertus* differs somewhat the Barents Sea assemblage from its counterpart in the Pechora region, and palynoflora from the Svalis-1 Formation (Vigran *et al.*, 1998; Mørk *et al.*, 1999) has been regarded therewith as younger than the early Griesbachian assemblage "P" (Hochuli *et al.*, 1989; Mangerud, 1994).

The *E. permasensis–P. disertus* assemblage of the Moscow region includes as well some species in common with the younger assemblage "O" from the upper part of the Vardebukta Formation (Hochuli *et al.*, 1989), analogues of which are established, despite the unsatisfactory preservation of miospores, in the Urd Formation (Mørk *et al.*, 1990) and in upper horizons of the Deltadalen Member in Spitsbergen (Mørk *et al.*, 1999). In addition to pollen forms *Lunatisporites* (*L. noviaulensis, L. pellucidus, L. transversundatus*), *Cycadopites*, and *P. pocockii*, the taxa in common include in this case the cavate spores *Densoisporites* spp., *Kraeuselisporites, Aratrisporites*, and *Pechorosporites* spp. Representatives of the genus *Pechorosporites* are more diverse in the assemblage "O" and include, in addition to *P. disertus* and *P. coronatus* present in the Moscow assemblage, the species *P. intermedius* Yaroshenko et Golubeva. The last feature implies that the assemblage "O" of the Barents Sea region is also close to the *P. disertus* assemblage of the Pechora syneclise, where spores of that genus are diverse and abundant. As compared to the concurrent palynoflora of the Moscow region, the assemblage "O" is much more rich in remains of planktonic organisms and lacks species *E. permasensis.*

Age of the topmost portion of the Vardebukta Formation of Spitsbergen is defined based on occurrence of *Vavilovites* aff. *sverdrupi* Tozer that is characteristic of the upper ammonite zone in the Dienerian Stage of Canada (Mørk *et al.*, 1999) or of the Induan Stage top. Based on comparison with the *D. nejburgii–P. pocockii–Maculatasporites* assemblage from the upper Griesbachian deposits bearing ammonites of the *Z. commune* Zone (Vigran *et al.*, 1998), as well as with the younger assemblage "O" from the Dienerian strata (Hochuli *et al.*, 1989; Mørk *et al.*, 1999), we may conclude that the *E. permasensis–P. disertus* assemblage of the Moscow syneclise and *P. disertus* assemblage of the Pechora syneclise are close in age to assemblages of the Barents Sea.

The materials considered above lead to the conclusion that the lower part of the Krasnye Baki Subformation (Sarafanikha Member) corresponds to the upper Griesbachian (*commune* Zone), although its younger age is also admissible. The upper Anisimovo Member barren of palynomorphs may be of the late Induan or early Olenekian age. It is overlain by the Shilikha or Rybinsk formations in the east and west of the Moscow syneclise, respectively. The latter formation yields palynomorphs of the Olenekian age (Yaroshenko, Lozovsky, 2002).

Table 2. Regional palynostratigraphy of marine and continental deposits of the Induan Stage as correlated with subdivisions of the general stratigraphic scale

Moscow syneclise				Barents Sea		Arctic Canada	East Greenland	Pechora syneclise	Chernyshev Chain of Hills	North Caspian syneclise	West European basin		Western Canada	General stratigraphic scale	
											Poland	Germany			
1				2		3	4	5	6	7	8	9	10	11	12
Formation	Subformation	Member	Palynological assemblage	Palynozone		Palynozone	Palynozone	Palynoflora	Palynoflora	Palynoflora	Palynozone	Palynoflora	Palynoflora	Stage	Ammonite zonation
Vokhma	Upper (Krasnye Baki)	Anisimovo	barren of palynomorphs	"O"				*Pechorosporites disertus*						Induan / Dienerian	*Vavilovites sverdrupi–Proptychites candidus*
		Sarafanikha	*Ephedripites permasensis-Pechorosporites disertus*	Svalis-1	*D. nejburgii–P. pocockii–Maculatasporites*									Griesbachian	*Bukkenites strigatus–Ophiceras commune*
	Lower	Ryabi	*Densoisporites complicatus–Ephedripites* sp.	"P"	*L. obsoleta–T. stoschiana*		*Lunatisporites (Taeniaesporites)*		*D. complicatus, Kraeuselisporites* spp., *Ephedripites* spp., *Lunatisporites* spp.	*Ephedripites* spp., *K. schaubergeri, T. stoschiana, L. novialensis*	*L. obsoleta–P. pantii*	*Endosporites papillatus–D. playfordi*	*Kraeuselisporites* spp., *U. imperialis, P. pocockii, L. jonkeri, N. striata, Ephedripites* spp., *Striatites* spp., *Lunatisporites* spp., *Alga*		*Otoceras boreale*
		Astashikha	*Striatoabieites richteri–Klausipollenites schaubergeri*			*T. stoschiana–S. richteri*	*Photohaploxypinus*		*Kraeuselisporites* spp., *Klausipollenities schaubergeri, L. virkkiae, Striatites* spp., *Lunatisporites* spp.						*Otoceras concavum*
		Nedubrovo	*Cycadopites* sp.–*K. schaubergeri*												

References for columns: (1) Lozovsky *et al.*, 2001a, 2001b; Yaroshenko and Goman'kov, 1998; Kyuntsel', 1965, 1969; Goman'kov *et al.*, 1986; Krassilov *et al.*, 1999a; Afonin, 2000; (2) Hochuli *et al.*, 1989; Mangerud, 1994; Vigran *et al.*, 1998; Mørk *et al.*, 1990; 1999; Fefilova, 1999; (3) Utting, 1994; (4) Balme, 1979; (5) Yaroshenko *et al.*, 1991; (6) Tuzhikova, 1985; (7) Makarova and Romanovskaya, 1983, 1999; (8) Orlowska-Zwolinska, 1984; Fijalkowska, 1994, 1995; Fijalkowska-Mader, 1998, 1999; (9) Reitz, 1985; Schultz, 1998, written communication; (10) Jansonius, 1962; (11) Tozer, 1967; (12) Orchard and Tozer, 1997.

DISCUSSION

Correlation based on palynological assemblages from continental and marine deposits (Table 2) is consistent in general with conclusions drawn from occurrence of amphibian taxon *Tupilakosaurus* in the Vokhma Horizon of the Moscow syneclise and in coastal-marine deposits of the Induan Stage of eastern Greenland (Shishkin *et al.*, 2000). Remains of the aforementioned amphibian genus, which are dispersed throughout the interval of the Vokhma Horizon, occur in eastern Greenland in association with ammonites of the *Glyptophiceras martini* and *Proptychites rosenkrantzi* local zones spanning the greater part of the lower Induan Substage in the interval of *boreale–strigatus* zones (Shevyrev, 2001). Palynological data add new details to this general correlation. The *Cycadopites* sp.–*K. schaubergeri* and *S. richteri–K. schaubergeri* palynological assemblages suggest that the Nedubrovo and basal Astashikha members span the interval of the ammonite *concavum* Zone, whereas the rest of the Astashikha interval and Ryabi Member can be attributed to the *boreale* Zone, as it follows from the range of *D. complicatus–Ephedripites* sp. assemblage characteristic of two aforementioned stratigraphic units. Synchronism of the *E. permasensis–P. disertus* assemblage from the lower Krasnye Baki Subformation and of palynoflora from the *D. nejburgii–P. pocockii–Maculatasporites* sp. palynozone of the Barents Sea, which yields the late Griesbachian ammonites, is inferable from palynological similarity between them. The concurrent ages of corresponding host deposits were established first more than 20 years ago based on remains of amphibians *Luzocephalus* found near the base of the Krasnye Baki Subformation and in deposits of the *Ophiceras commune* Zone of eastern Greenland (Shishkin, 1980). In the Moscow syneclise, this zone separates the lower and upper subformations of the Vokhma Formation and represents one of the most distinct boundary horizons of regional significance. According to paleomagnetic data, it marks the boundary between the normal (N1T) and reversed (R1') polarity zones. In the Sverdrup basin of Canada, the same polarity reversal is recorded at the *boreale/commune* interface in the ammonite scale (Ogg and Steiner, 1989),[2] and the inferred correlation between palynological and ammonite zones seems to be valid. An older reversed polarity zone (R0T) established by B.V. Burov in the Nedubrovo Member (Lozovsky *et al.*, 2001) is also traceable in other Triassic sections, e.g., near the base of the Deltadalen Member (Mørk *et al.*, 1999).

[2] In sections of Spitsbergen (Mørk *et al.*, 1999), this paleomagnetic event is recorded slightly higher, inside the *commune* Zone. The discrepancy is minor and can be removed after additional paleomagnetic and biostratigraphic study.

CONCLUSION

The studied palynological assemblages characterizing successive members of the Vokhma Formation in the Moscow syneclise have many features in common despite all the variations in abundance and diversity of their taxa. Consequently, the Early Triassic palynofloras developed gradually succeeding older assemblages. Yaroshenko and Goman'kov (1998) noted this trend beginning from the lower *Tympanicysta stoschiana–Ephedripites* spp. assemblage of the Moloma Member at the Vyatka Horizon top, where initial changes are recorded in composition of palynofloras bearing the inherited character.

Demonstrating the growing abundance of Triassic components, palynological assemblages of the Vokhma Formation still include some taxa, which are typical of the Upper Permian and impart an older aspect to palynofloras. A considerable quantity of Permian taxa is characteristic as well of the earliest Triassic palynofloras in the other regions of the world (Balme, 1979; Orlowska-Zwolinska, 1984; Hochuli *et al.*, 1989; Ouyang and Utting, 1990; Mangerud, 1994; Utting, 1994).

It is noteworthy that all the studied and considered assemblages of microflora include species *Tympanicysta stoschiana* of green algae, which are of a worldwide distribution in continental and marine deposits of the Upper Permian and Lower Triassic. The suggested succession of palynological assemblages in the Lower Triassic continental sequence of the Moscow syneclise needs in confirmations based on additional palynological analysis of different sedimentary facies. The established *Cycadopites* sp.–*Klausipollenites schaubergeri* and *Striatoabieites richteri–K. schaubergeri* palynological assemblages are correlated with the ammonite *Otoceras concavum* Zone, whereas younger *Densoisporites complicatus–Ephedripites* sp. and *Ephedripites permasensis–Pechorosporites disertus* assemblages are attributed to the *Otoceras boreale* and *Ophiceras commune* zones, respectively. The last youngest assemblage probably spans in addition some higher interval of the Induan Stage. The suggested correlation is consistent with data obtained by ecotone and paleomagnetic methods. In view of the results obtained, the idea of V.V. Menner who thought that palynological method is applicable for the direct correlation of continental and marine facies seems to be the rightful one.

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Radiolarians from Upper Cretaceous Deposits, the Novodevich'e Section (Samara Oblast, Volga River Middle Courses)

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Abstract—Radiolarians from the Novodevich'e section, Samara oblast, are described. Two subdivisions of the section correspond to the 6.5-m-thick member of clay and chalk with Turonian–Coniacian benthic foraminifers and to the other member of intercalated opoka and clay beds, which are 6.8 m thick in total and yield benthic foraminifers of the upper Coniacian–Santonian in association with radiolarians. Of 41 radiolarian species, which have been identified, the most typical are *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *C. cachensis* Pessagno, *C. latum* (Lipman), *Orbiculiforma monticelloensis* Pessagno, *O. quadrata* Pessagno, *Paronaella santonica* (Lipman), *P. tumida* (Lipman), *Patulibracchium ingens* (Lipman), *Pentinastrum subbotinae* Lipman, *Praeconocaryomma lipmanae* Pessagno, *P. universa* Pessagno, *Pseudoaulophacus lenticulatus* (White), *Triactoma compressa* (Squinabol), *Amphipyndax stocki* (Campbell et Clark), *Dictyomitra multicostata* Zittel, and *Xitus asymbatos* (Foreman). The late Coniacian–Santonian age of radiolarian assemblage is inferred based on coexisting radiolarians and on correlation with radiolarians characteristic of the Zagorsk Formation of the Moscow syneclise. As compared to radiolarian assemblages of the Moscow syneclise, which are taxonomically diverse and include many taxa known from California and bottom sediments of tropical oceanic regions, and to concurrent assemblages of the Urals and West Siberia, which are depleted in thermophilic species, the Late Cretaceous radiolarians from the Volga River middle courses are of a transitional taxonomic composition. It is plausible to conclude therefore that thermophilic taxa migrated into the East European sea from the west, whereas cryophilic forms characteristic of the West Siberian basin arrived from the east and northeast. Species originally identified by Lipman (1952) are revised and described anew.

Key words: Upper Cretaceous, stratigraphy, radiolarians, foraminifers, Volga River basin.

INTRODUCTION

Radiolarians from Upper Cretaceous (Santonian–Maastrichtian) deposits in middle courses of the Volga River have been described first by Lipman (1952) who studied core sections recovered by drilling in the east of the Penza oblast. Later on, they have been investigated in other sections of the region located near the Ul'yanovsk (Bragina, 1987), Saratov (Kazintseva, 2001), and between the Volga and Don rivers (Bragina *et al.*, 1999). These microfossils are of particular interest for the detailed subdivision and correlation of Upper Cretaceous (predominantly Santonian–Campanian) siliceous deposits, which, being widespread in the study region, rarely yield inoceramids and foraminifers, the main orthostratigraphic fossils of the Upper Cretaceous. Moreover, the Late Cretaceous radiolarians of the region essentially differ in diversity and taxonomic composition from radiolarian assemblages in bottom sediments of the World Ocean tropical zone, which have been used to elaborate the Upper Cretaceous radiolarian zonation, the only one that is sufficiently detailed and complete at present (Sanfilippo and Reidel, 1985). In addition, the Late Cretaceous radiolarians from the region under consideration are inadequately studied: they are lacking classical description and stratigraphic succession of their assemblages has not been analyzed. Accordingly, it is difficult to assess stratigraphic potential, paleofacies and paleoclimatic significance of this fossil group. Therefore, investigation of radiolarians is an actual problem for the Upper Cretaceous reference sections in middle courses of the Volga River.

METHODS

Radiolarians have been studies in clay samples (about 200 g in weight), which are more suitable for radiolarian analysis than opokas, in which radiolarian skeletons are significantly recrystallized because of the late diagenetic redistribution of silica. The collected samples have been disintegrated first in hydrogen peroxide (H_2O_2) and treated then in diluted acetic acid (CH_3COOH) in order to eliminate carbonate particles (remains of inoceramid prismatic layer). Radiolarians have been picked up from dry residues under binocular microscope and photographed under the scanning electron microscope Cambridge Stereoscan 600 at the Geological Institute, Russian Academy of Sciences. We are grateful to N.V. Gor'kova who performed the electron-microscopic research.

SECTION AND STRATIGRAPHIC SUCCESSION OF MICROFAUNAS

The sedimentary sequence studied near the Novodevich'e Village, Samara oblast (Fig. 1), have been selected earlier as an object for excursions of the International Geological Congress of 1984 (Adas *et al.*, 1984). Rocks of the sequence are perfectly exposed and yield abundant well-preserved radiolarians in association with informative benthic foraminifers (preliminary identified by V.N. Beniamovskii). Sediments for radiolarian analysis have been sampled in 1989 by N.Yu. Bragin, and E.Yu. Baraboshkin collected additional samples bearing foraminifers in 2000. The section description presented below is based on field observations of 1989 (Fig. 2).

The Albian deposits at the base of the section are represented by dark gray to black silty clay with rare black phosphorite nodules. The overlying Upper Cretaceous sequence is composed of the following beds.

Bed 1: greenish black medium-grained glauconite sand with redeposited pebbles of Albian phosphorite; sand discordantly overlies the eroded surface of Albian clay (0.2 to 0.4 m thick);

Bed 2: light gray calcareous–sandy clay displaying gradual transitions to underlying and overlying sediments (0.2 m thick); beds 1 and 2 yield foraminifers of the middle Turonian *Gavelinella moniliformis* Zone (Akimets *et al.*, 1991);

Bed 3: grayish compact sandy chalk with blocky jointing (about 6 m thick); in the interval from 0 to 2 m above the bed base, there is established a succession of the upper Turonian *Gavelinella moniliformis* and *Gavelinella praeinfrasantonica* foraminiferal zones (Akimets *et al.*, 1991). The level of 2 m above the base corresponds to an erosion surface, above which foraminifers of the upper Coniacian *Gavelinella thalmanni* Zone and index subspecies *Stensioeina exculpta exculpta* of the upper Coniacian–lower Santonian have been encountered (Akimets *et al.*, 1991);

Bed 4: light gray sandy opoka; the bed base corresponds to an erosion surface with phosphorite and chalk pebbles (0.3 m thick);

Bed 5: light colored, greenish gray marly clay; the bed is flaggy (0.4 m thick);

Bed 6: massive opoka of the same coloration; the bed is 0.6 m thick;

Bed 7: clay like that of Bed 5 (0.5 m thick); Sample 21 from the bed yielded radiolarians *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *C. irwini* Pessagno, *C. latum* (Lipman), *Dispongotripus triangularis* (Squinabol), *Orbiculiforma impressa* (Lipman), *O. quadrata* Pessagno, *O. persenex* Pessagno, *O. vacaensis* Pessagno, *Paronaella tumida* (Lipman), *Pentinastrum subbotinae* Lipman, *Pseudoaulophacus praefloresensis* Pessagno, *Triactoma compressa* (Squinabol), *Vitorfus brustolensis* (Squinabol), and *Dictyomitra densicostata* Pessagno;

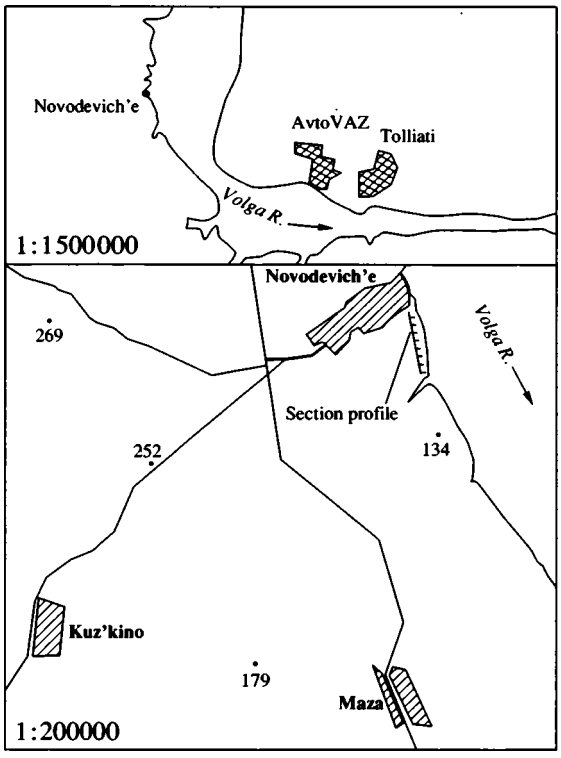


Fig. 1. Locality of Upper Cretaceous deposits near the Novodevich'e Village (figures denote altitudes in meters).

Bed 8: opoka like that of Bed 6 (1 m thick); beds 4 to 8 bear foraminifer species *Neoflabellina suturalis* of the upper Coniacian–lower Santonian (Akimets *et al.*, 1991);

Bed 9: clay like that of Bed 5 (0.5 m thick); radiolarians identified in Sample 22 from this bed are *Acaeniotyle umbilicata* (Rust), *Archaeospongoprunum bipartitum* Pessagno, *A. cortinaensis* Pessagno, *A. hueyi* Pessagno, *Crucella aster* (Lipman), *C. cachensis* Pessagno, *C. latum* (Lipman), *C. messinae* Pessagno, *Cromyodruppa concentrica* Lipman, *Orbiculiforma monticelloensis* Pessagno, *Patulibracchium* aff. *P. petroleumensis* Pessagno, *P. ingens* (Lipman), *Paronaella santonica* (Lipman), *P. tumida* (Lipman), *Pentinastrum subbotinae* Lipman, *Phaseliforma* sp. cf. *P. subcarinata* Pessagno, *Praeconocaryomma californiaensis* Pessagno, *P. lipmanae* Pessagno, *P. universa* Pessagno, *Pseudoaulophacus parguerensis* Pessagno, *Spongopyle insolita* group Kozlova, *Amphipyndax stocki* (Campbell et Clark), *Archaeodictyomitra* sp. ex gr. *A. squinaboli* Pessagno, *Dictyomitra densicostata* Pessagno, *D. multicostata* (Zittel), *Stichomitra cechena* Foreman, and *Xitus asymbatos* (Foreman).

Bed 10: opoka like that of Bed 6; the bed is 0.6 m thick;

Bed 11: clay like that of Bed 5 (0.3 m thick), radiolarians identified in Sample 23 from the bed are *Acaeniotyle umbilicata* (Rust), *Crucella aster* (Lipman),

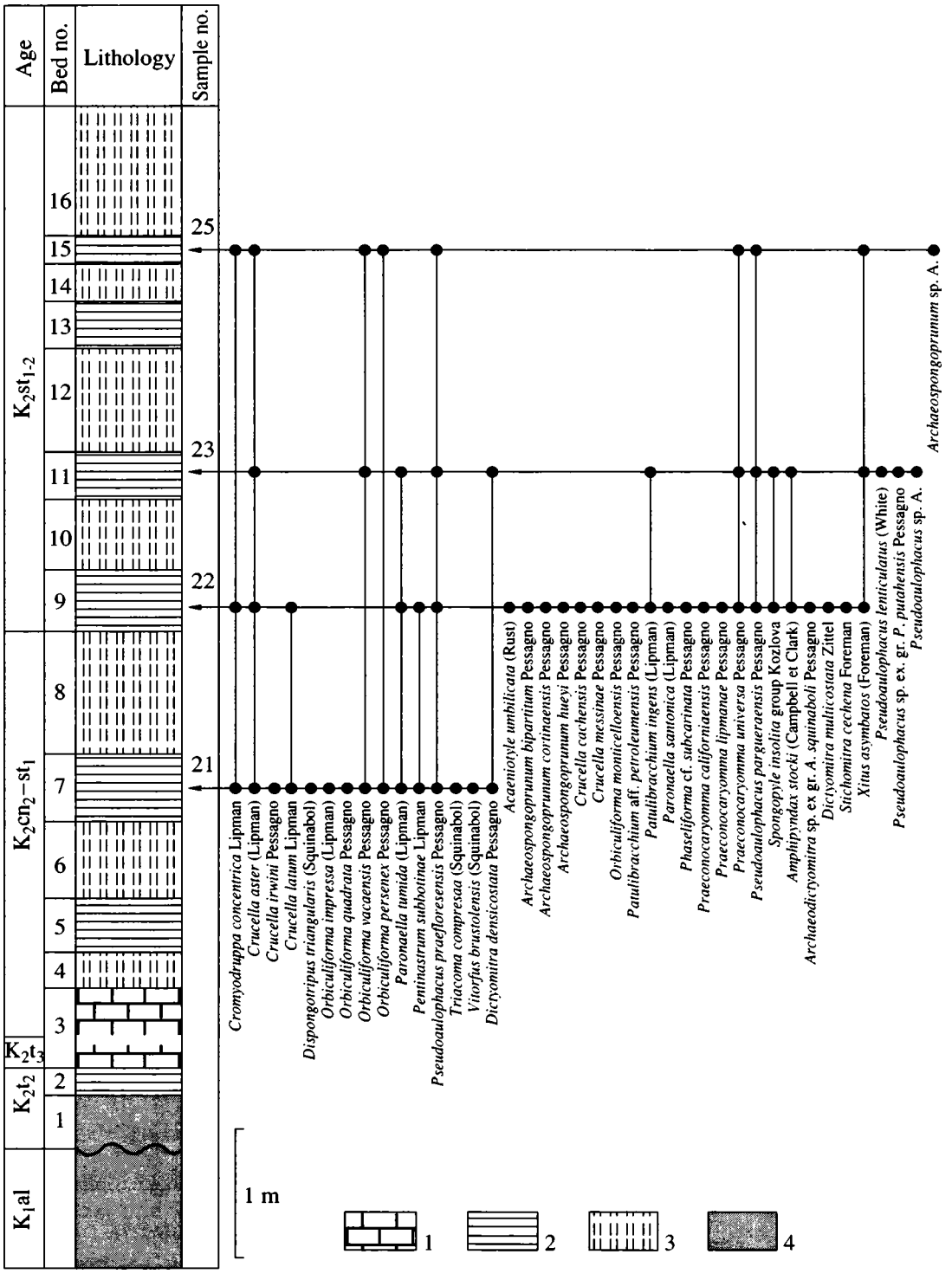


Fig. 2. Stratigraphic ranges of radiolarians in the Novodevich'e section of Upper Cretaceous deposits (Bed 3 is shown out of scale): (1) chalk-like limestone; (2) clay; (3) opoka; (4) sand.

Orbiculiforma vacaensis Pessagno, *Paronaella tumida* (Lipman), *Patulibracchium ingens* (Lipman), *Praeconocaryomma universe* Pessagno, *Pseudoaulophacus lenticulatus* (White), *Pseudoaulophacus parqueraensis* Pessagno, *Ps. praefloresensis* Pessagno, *Ps.* sp. ex gr. *Ps. putahensis* Pessagno, *Ps.* sp. A, *Spongopyle insolita* group Kozlova, *Amphipyndax stocki* (Campbell et Clark), *Dictyomitra densicostata* Pessagno, and *Xitus asymbatos* (Foreman).

Bed 12: opoka like that of Bed 6 (0.8 m thick);

Bed 13: clay like that of Bed 5 (0.3 m thick);

Bed 14: opoka like that of Bed 6 (0.3 m thick);

Bed 15: clay like that of Bed 5 (0.2 m thick); the bed (Sample 25) yields radiolarians *Archaeospongoprunum* sp. A, *Alievium* sp. A, *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *Orbiculiforma persenex* Pessagno, *O. vacaensis* Pessagno, *Praeconocaryomma universa* Pessagno, *Pseudoaulophacus parguerensis* Pessagno, *P. praefloresensis* Pessagno, and *Xitus asymbatos* (Foreman). In addition, beds 9–15 yield foraminifers *Stensioeina exculpta gracilis* of the terminal lower–upper Santonian (Akimets *et al.*, 1991).

Bed 16: opoka like that of Bed 6, the apparent thickness is 1 m.

After the next unexposed interval, there are low hills with detritus of gray and white quartz sandstones (Paleogene?). The presumable thickness of sandstones is 3 to 4 m.

RESULTS AND DISCUSSION

Radiolarians found at several level within the upper Coniacian–Santonian interval of the sequence represent in fact one diverse assemblage (Fig. 2) consisting of 41 species. These are *Acaeniotyle umbilicata* (Rust), *Alievium* sp. A, *Archaeospongoprunum bipartitum* Pessagno, *A. cortinaensis* Pessagno, *A. hueyi* Pessagno, *Archaeospongoprunum* sp. A, *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *C. cachensis* Pessagno, *C. irwini* Pessagno, *C. latum* (Lipman), *C. messinae* Pessagno, *Dispongotripus triangularis* (Squinabol), *Orbiculiforma impressa* (Lipman), *O. monticelloensis* Pessagno, *O. quadrata* Pessagno, *O. persenex* Pessagno, *O. vacaensis* Pessagno, *Paronaella santonica* (Lipman), *P. tumida* (Lipman), *Patulibracchium ingens* (Lipman), *P.* aff. *P. petroleumensis* Pessagno, *Pentinastrum subbotinae* Lipman, *Phaseliforma* sp. cf. *P. subcarinata* Pessagno, *Praeconocaryomma californiaensis* Pessagno, *Pr. lipmanae* Pessagno, *Pr. universa* Pessagno, *Pseudoaulophacus lenticulatus* (White), *Ps. parguerensis* Pessagno, *Ps. praefloresensis* Pessagno, *Ps.* sp. ex gr. *Ps. putahensis* Pessagno, *Ps.* sp. A, *Spongopyle insolita* group Kozlova, *Triactoma compressa* (Squinabol), *Vitorfus brustolensis* (Squinabol), *Amphipyndax stocki* (Campbell et Clark), *Archaeodictyomitra* sp. ex gr. *A. squinaboli* Pessagno, *Dictyomitra densicostata* Pessagno, *D. multicostata* Zittel, *Stichomitra cechena* Foreman, and *Xitus asymbatos* (Foreman). Radiolarians are most diverse in Sample 22 (Bed 9), but all the species from this level are of a broad stratigraphic range.

Fifteen of the listed species have been originally described in California (Campbell and Clark, 1944; Pessagno, 1971, 1972, 1976), and many of them are of a wide geographic range. *Archaeospongoprunum bipartitum* Pessagno and *Orbiculiforma quadrata* Pessagno are known from Coniacian deposits of California (Pessagno, 1976), but in the Russian platform (e.g., in Moscow region) they are common in Santonian beds as well (Bragina, 1994). Species *Archaeospongoprunum cortinaensis* Pessagno, *Triactoma compressa* (Squinabol), and *Vitorfus brustolensis* (Squinabol) are characteristic of Cenomanian and Turonian sediments of the Tethyan region (Bragina, 1999; 2001; O'Dogherty, 1994; Salvini and Marcucci Passerini, 1998), although we do not exclude their occurrence in younger deposits, because the Coniacian–Santonian interval is insufficiently studied in Mediterranean areas. Such species as *Acaeniotyle umbilicata* (Rust), *Crucella latum* (Lipman), *Pseudoaulophacus lenticulatus* (White), *Amphipyndax stocki* (Campbell et Clark), *Dictyomitra multicostata* Zittel, and *Xitus asymbatos* (Foreman) occur worldwide in Cenomanian–Campanian strata. *Cromyodruppa concentrica* Lipman, *Paronaella santonica* (Lipman), *P. tumida* (Lipman), *Patulibracchium ingens* (Lipman), and *Pentinastrum subbotinae* Lipman are widespread in Coniacian–Campanian deposits of the Russian platform (Bragina, 1994; Bragina *et al.*, 1999; Vishnevskaya, 1987; Lipman, 1952; Olfer'ev *et al.*, 2000) and West Siberia (Kozlova and Gorbovets, 1966). The above data on distribution of radiolarians suggest the Coniacian–Santonian age of the assemblage, but coexisting foraminifers restrict it by the late Coniacian–Santonian time span.

It is interesting to compare the established assemblage with concurrent radiolarian faunas from Saratov and Ul'yanovsk areas of the Volga River basin. Kazintsova (2001) distinguished previously the early and late Santonian assemblages of radiolarians. Our assemblage includes six species characteristic of her early Santonian assemblage: *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *Paronaella tumida* (Lipman), *Pseudoaulophacus lenticulatus* (White), *Amphipyndax stocki* (Campbell et Clark), and *Dictyomitra densicostata* Pessagno. Fourteen taxa from our assemblage represent species in common with her late Santonian assemblage. These are *Archaeospongoprunum bipartitum* Pessagno, *A. cortinaensis* Pessagno, *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *C. latum* (Lipman), *Orbiculiforma monticelloensis* Pessagno, *O. persenex* Pessagno, *O. vacaensis* Pessagno, *Paronaella santonica* (Lipman), *P. tumida* (Lipman), *Pseudoaulophacus lenticulatus* (White), *Amphipyndax stocki* (Campbell et Clark), *Dictyomitra densicostata* Pessagno, and *D. multicostata* Zittel. The concurrent assemblage of the Ul'yanovsk area (Bragina, 1987) includes nearly all the species of the Samara assemblage except for *Dispongotripus triangularis* (Squinabol), *Phaseliforma* sp. cf. *P. subcarinata* Pessagno, *Ps.* ex gr. *Ps. putahensis* Pessagno, *Ps.* sp. A, *Triactoma compressa* (Squinabol), and *Vitorfus brustolensis* (Squinabol). Accordingly it is safe to conclude that the late Coniacian–Santonian radiolarian assemblages from different areas in middle courses of the Volga River are close to each other in taxonomic composition.

A comparison with concurrent or nearly concurrent assemblages of the Moscow region (Bragina, 1994) and Vladimir area (Olfer'ev *et al.*, 2000) is also interesting.

The assemblage from the Zagorsk Formation of the Moscow region includes 22 species in common with the assemblage of the Novodevich'e section. These are *Archaeospongoprunum bipartitum* Pessagno, *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *C. irwini* Pessagno, *C. latum* (Lipman), *Orbiculiforma quadrata* Pessagno, *O. persenex* Pessagno, *O. vacaensis* Pessagno, *Paronaella santonica* (Lipman), *P. tumida* (Lipman), *Patulibracchium ingens* (Lipman), *Pentinastrum subbotinae* Lipman, *Praeconocaryomma californiaensis* Pessagno, *Pf. lipmanae* Pessagno, *Pr. universa* Pessagno, *Pseudoaulophacus praefloresensis* Pessagno, *Spongopyle insolita* group Kozlova, *Amphipyndax stocki* (Campbell et Clark), *Dictyomitra densicostata* Pessagno, *D. multicostata* Zittel, *Stichomitra cechena* Foreman, and *Xitus asymbatos* (Foreman). It is remarkable therewith that the assemblage from Moscow region is lacking two species of a wide geographic range: *Archaeospongoprunum cortinaensis* Pessagno and *Vitorfus brustolensis* (Squinabol). On the other hand, *Pseudoaulophacus floresensis* Pessagno, *Stylosphaera pusilla* Campbell et Clark, *Cryptamphorella conara* (Foreman), and *Tricolocapsa granti* Campbell et Clark, which are the widespread Late Cretaceous species known in Moscow region, have not been encountered in the Novodevich'e section. In general, the assemblage of Moscow region is more diverse and includes a greater number of species, which are characteristic of warm-water basins.

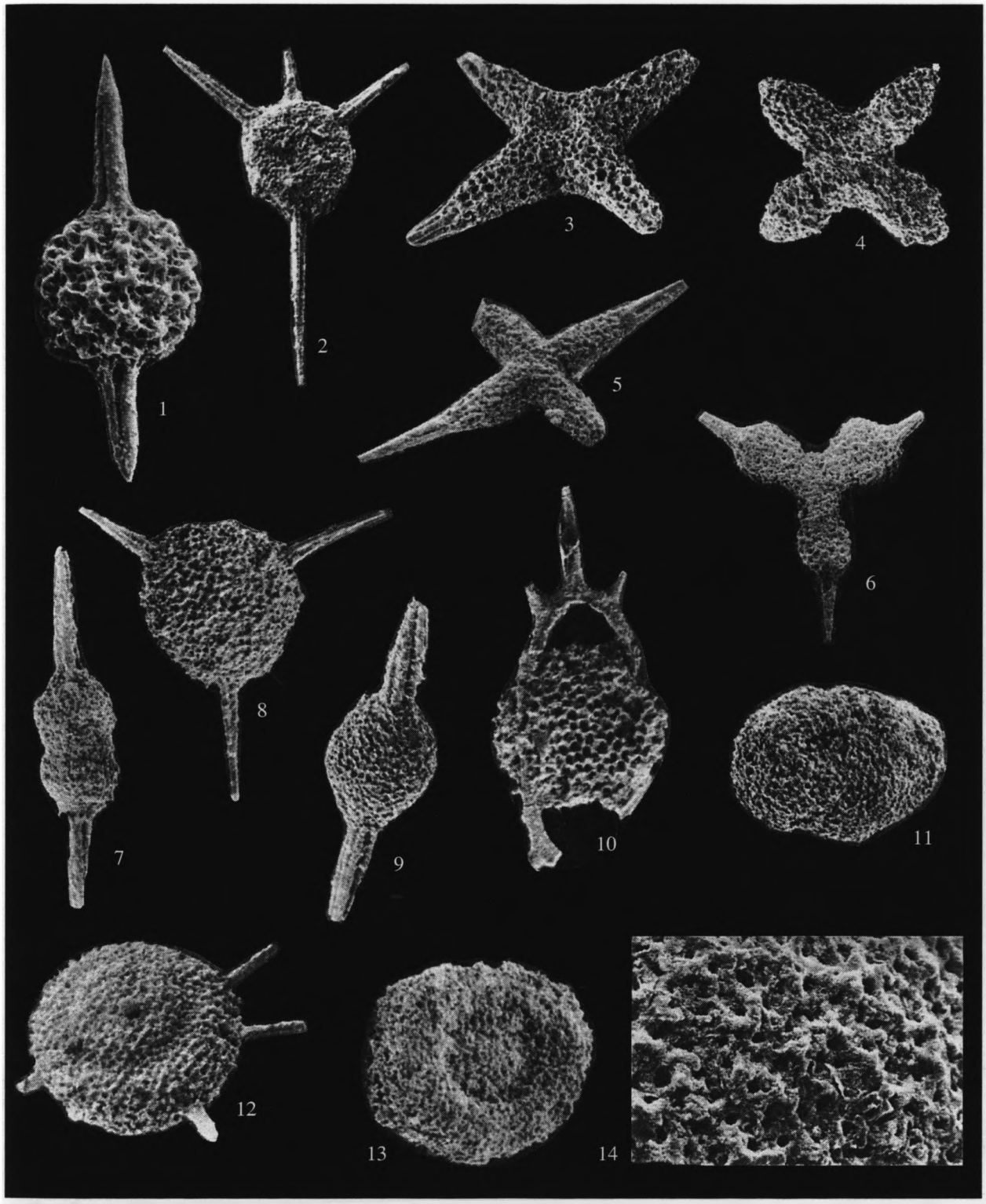
Comparing the assemblages of the Vladimir area and Novodevich'e section, we established nine species in common. These are *Archaeospongoprunum bipartitum* Pessagno, *Cromyodruppa concentrica* Lipman, *Orbiculiforma quadrata* Pessagno, *O. persenex* Pessagno, *O. vacaensis* Pessagno, *Paronaella tumida* Lipman, *Crucella aster* (Lipman), *C. latum* (Lipman), and *Xitus asymbatos* (Foreman). We suspect that the assemblage of Vladimir area just appears to be less diverse, because radiolarians are preserved here not as well as in the Moscow region.

Seven species in common established by comparison with Santonian assemblages from southern areas east of the Urals are as follows: *Archaeospongoprunum bipartitum* Pessagno, *Cromyodruppa concentrica* Lipman, *Crucella aster* (Lipman), *C. latum* (Lipman), *Praeconocaryomma universa* Pessagno, *Pseudoaulophacus lenticulatus* (White), and *Dictyomitra multicostata* Zittel. Assemblages from areas situated westward of the sub-Polar and northern Urals (Amon, 2000) include only two species occurring in the studied section (*Amphipyndax stocki* (Campbell et Clark) and *Dictyomitra multicostata* Zittel). It is remarkable as well that several species characteristic in general of the West Siberian radiolarians are missing from the assemblage of the Novodevich'e section. These are *Crucella tumeniensis* Lipman, *Porodiscus vulgaris* Lipman, *Spongotripus morenoensis* Campbell et Clark, *Stylotrochus dolichacanthus* (Lipman), and *Theocampe animula* (Kozlova and Gorbovets, 1966).

CONCLUSION

The late Coniacian–Santonian radiolarian assemblage of the Novodevich'e section is well correlative with concurrent assemblages of the Ul'yanovsk, Moscow, and Vladimir areas. Distinctions between these assemblages are suitable for the subsequent paleoclimatic and paleogeographic reconstructions. For instance, it is clear that the studied assemblage is transitional in taxonomic composition, connecting the assemblages of the Moscow syneclise, which are taxonomically diverse (Bragina, 1994), including many taxa known from California (Pessagno, 1976) and from bottom sediments of tropical oceanic regions (Sanfilippo and Riedel, 1985), and of the Urals and West Siberia, which are depleted in thermophilic species. Factors responsible for this can be found in peculiarities of the Late Cretaceous epicontinental seas of Eastern Europe and West Siberia and in their connections between each other and with adjacent sea basins. The West Siberian sea was undoubtedly well connected with the Arctic Ocean at that time but weakly communicated with basins of the Turan plate. This situation determined specific taxonomic compositions of radiolarian assemblages in West Siberia and nearby areas of the Urals, which are regarded as characteristic of the Boreal realm (Amon, 2000; Vishnevskaya, 2001). The East European basin could be connected with the West Siberian sea via straits that crossed the central Urals and Pai-Khoi Ridge (Amon and De Wever, 1994). Via a narrow meridional strait along the western flank of the Urals, it could be connected with Arctic cold waters, which penetrated to the middle stream area of the Volga River. Connections between the Tethys and East and West European seas were broad, favorable for ingression of warm waters into southern and western areas of the latter. Such an ingression was likely responsible for the relatively warm-water affinity of radiolarian assemblages of the Moscow syneclise and for a transitional status of radiolarians characterizing sections in middle courses of the Volga River.

Some species of Late Cretaceous radiolarians, which were described by Lipman as long ago as in the 1950s and currently need in revision, are described anew in the paleontological section of the work. Abilities of electronic microscopy have been used to improve morphological characterization of these species. Their systematics is specified with due account for all the semi-centennial changes in taxonomy of radiolarians.



Plates I–II. Late Coniacian–Santonian radiolarians from the Novodevich'e section, Samara oblast.

Plate I. (1) *Acaeniotyle umbilicata* (Rust), Sample 22, ×150; (2) *Dispongotripus triangularis* (Squinabol), Sample 21, ×80; (3) *Crucella cachensis* Pessagno, Sample 22, ×150; (4) *Crucella aster* (Lipman), Sample 21, ×150; (5) *Crucella messinae* Pessagno, Sample 22, ×150; (6) *Paronaella santonica* (Lipman), Sample 22, ×120; (7) *Archaeospongoprunum bipartitum* Pessagno, Sample 22, ×150; (8, 9) *Archaeospongoprunum cortinaensis* Pessagno, Sample 22, ×80 (8) and ×150 (9); (10) *Vitorfus brustolensis* (Squinabol), Sample 21, ×200; (11, 14) *Phaseliforma* cf. sp. *P. subcarinata* Pessagno, Sample 22, general view, ×150 (11) and perforation style of outer wall, ×800 (14); (12) *Pseudoaulophacus lenticulatus* (White), Sample 23, ×150; (13) *Orbiculiforma monticelloensis* Pessagno, Sample 22, ×150.

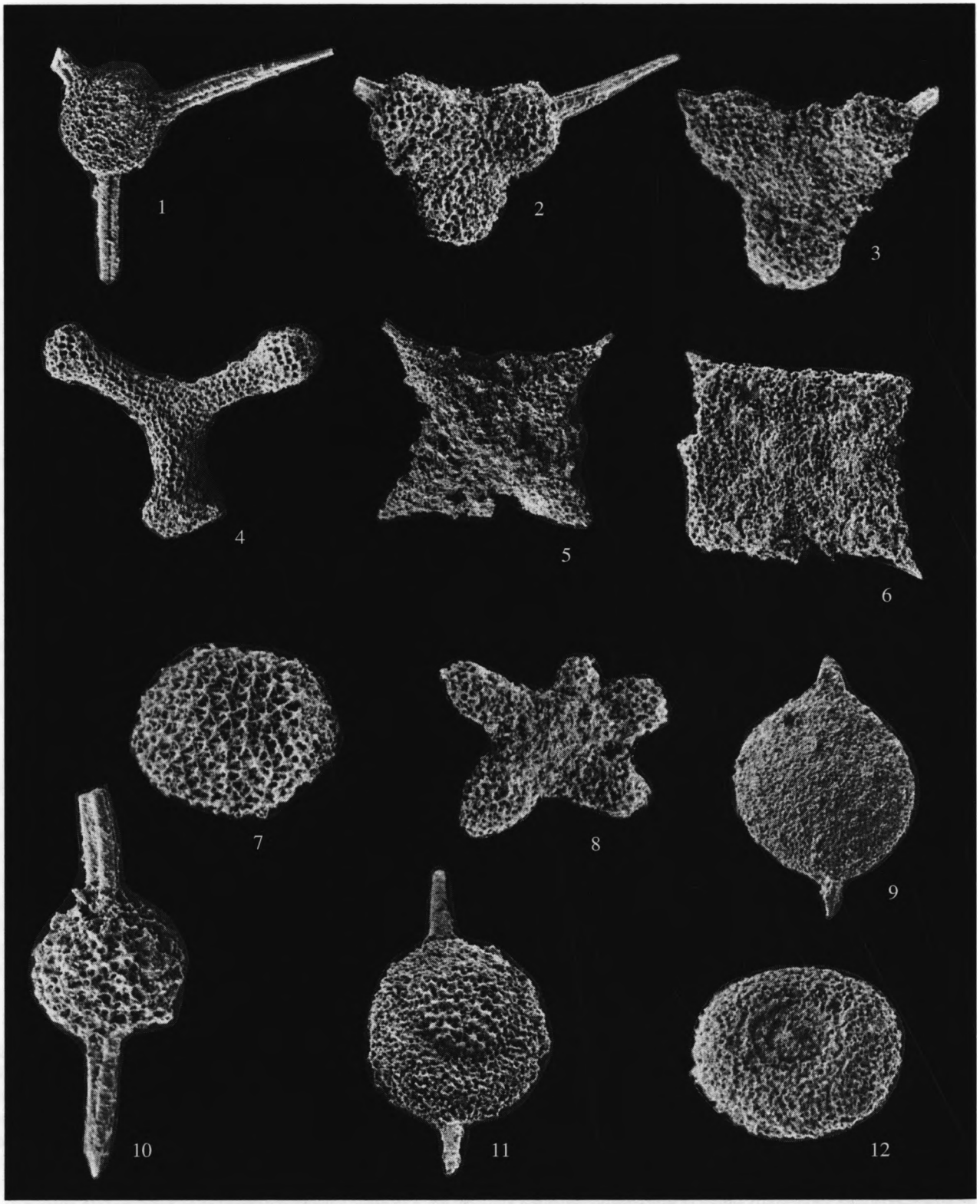


Plate II. (1) *Triactoma compressa* (Squinabol), Sample 21, ×80; (2, 3) *Paronaella tumida* (Lipman), Sample 21, ×150, (2) and ×140, (3); (4) *Patulibracchium ingens* (Lipman), Sample 22, ×65; (5, 6) *Crucella latum* (Lipman), from samples 21 (5) and 22 (6), ×100 in both figs.; (7) *Alievium* sp. A, Sample 25, ×150; (8) *Pentinastrum subbotinae* Lipman, Sample 22, ×100; (9) *Archaeospongoprunum* sp. A, Sample 25, ×150; (10) *Acaeniotyle umbilicata* (Rust), Sample 22, ×150; (11) *Pseudoaulophacus* sp. A, Sample 23, ×150; (12) *Orbiculiforma vacaensis* Pessagno, Sample 23, ×150.

PALEONTOLOGICAL DESCRIPTIONS

Family Hagiastridae Riedel, 1971

Subfamily Hagiastrinae Riedel, 1967, emend. Pessagno, 1971

Genus *Crucella* Pessagno, 1971

Crucella aster (Lipman, 1952)

Plate I, Fig. 4

Histiastrum aster: Lipman, 1952, p. 35, Plate II, figs. 6, 7; 1962, p. 300, Plate II, fig. 5; Kozlova and Gorbovets, 1966, p. 84, Plate III, fig. 9; Amon, 2000, p. 51, Plate VI, fig. 15; Vishnevskaya, 2001, Plate 115, fig. 6

Crucella aster: Kazintsova and Olfer'ev, 1997, Plate I, fig. 1 (nomen nudum).

Holotype: no.16/28, collection no. 6999, F.N. Chernyshev Central Geological Museum; Kuznetsk locality, Penza oblast, Santonian (Russian plate).

Description. Tests are medium-sized, with arms of quadrangular radiation. Central capsule of the test is not elevated. Each of the arms proximally broad is narrowing toward distal end that is terminated by a massive spine. The test surface is irregularly perforated by rounded-polygonal pores.

Dimensions. Two neighboring spine ends are 247 to 437 μm away from each other, spines are 152 to 57 μm long and 65 to 95 μm thick at maximum. Terminal needles are 20 to 60 μm long.

Remarks. Kazintsova and Olfer'ev (1997) attributed the species to the genus *Crucella*, but they did publish its paleontological description.

Distribution: worldwide, terminal upper Albian–Campanian.

Material: dozens of specimens.

Crucella latum (Lipman, 1960)

Plate II, figs. 5, 6

Histiastrum latum: Lipman *et al.*, 1960, p. 130, Plate XXIX, figs. 7, 8; Bragina, 1994, Fig. 1, fig. 9; Amon, 2000, p. 51, Plate VI, fig. 17; Vishnevskaya, 2001, p. 163, Plate 114, fig. 9.

Holotype: no. 56/3, collection no. 7767, Central Geological Museum, St. Petersburg (Plate XXIX, figs. 7, 8); West Siberia, lower part of the lower radiolarian sequence, Borehole 1-P; Santonian–Campanian.

Description. Medium-sized tests are quadrangular in plane. The test surface is irregularly perforated by rounded-polygonal pores. Apical spines are massive and long, usually round in cross-section.

Dimensions. Tests are 180 to 300 μm long, and needles are as long as 30–70 μm.

Comparison. Spines of irregular shape differ *Crucella latum* (Lipman, 1960) from *Crucella espartoensis* Pessagno (Pessagno, 1971).

Distribution: worldwide, terminal upper Albian–Campanian.

Material: dozens of specimens.

Genus *Paronaella* Pessagno, 1971

Paronaella santonica (Lipman, 1952)

Plate I, fig. 6

Euchitonia santonica: Lipman, 1952, p. 34, Plate II, fig. 3; Vishnevskaya, 2001, p. 162, Plate 114, fig. 11.

Holotype: no. 16/42, collection no. 6999, F.N. Chernyshev Central Geological Museum; the Kuznetsk site of Penza oblast, Santonian (Russian plate).

Description. Discoid skeletons are subtriangular in plane. Three short arms are facing vertices of equilateral (or isosceles) triangle. Being swollen, the arms are wider in plane than central region. Thick apical spines are frequently long, faceted near the base and conical at distal ends. Their length is 1.5–2 times greater than the thickest part. Arms are spongy inside. The internal part of central capsule is of annulate structure. External surface is irregularly perforated by minute pores of rounded polygonal shape. Patagium that is frequently preserved extends over arms concealing the spine basal ends.

Dimensions. Tests are 255–270 and 200–250 μm in diameter with and without spines, respectively; spines are 30 to 70 μm long.

Comparison. In distinction from *Paronaella venadoensis* Pessagno (Pessagno, 1971), tests have no by-spines on radial spines, their arms are more swollen and perforations are smaller in diameter.

Distribution: Coniacian–Campanian of the Russian plate and West Siberia.

Material: 24 specimens.

Paronaella tumida (Lipman, 1952)

Plate II, fig. 2, 3

Rhopalastrum tumidum: Lipman, 1952, p. 37, Plate II, fig. 13; Bragina, 1994, Fig. 1, fig. 11

Holotype: no. 16/43, collection no. 6999, F.N. Chernyshev Central Geological Museum; Kuznetsk locality of Penza oblast., Santonian (Russian plate).

Description. Large discoid skeletons are subtriangular in plane. Three arms radiating toward vertices of equilateral triangle are very short. Central region is equal in size to basal diameters of arms. Small central tholus is insignificantly elevated. Arms are wider than central region, considerably swollen and rounded in shape. Apical spines are thick, frequently long, faceted near their base and conical at distal ends. Their length is almost equal in size to the thickest part diameter. Arms are spongy inside. The internal part of central region is of indistinct annulate structure. External surface is irregularly perforated by minute close-spaced pores of rounded polygonal shape. Patagium is often developed over the test surface and spine basal ends.

Dimensions. Skeletons are 240–270 μm in diameter, arms are 120–150 μm long and 95–110 μm wide,

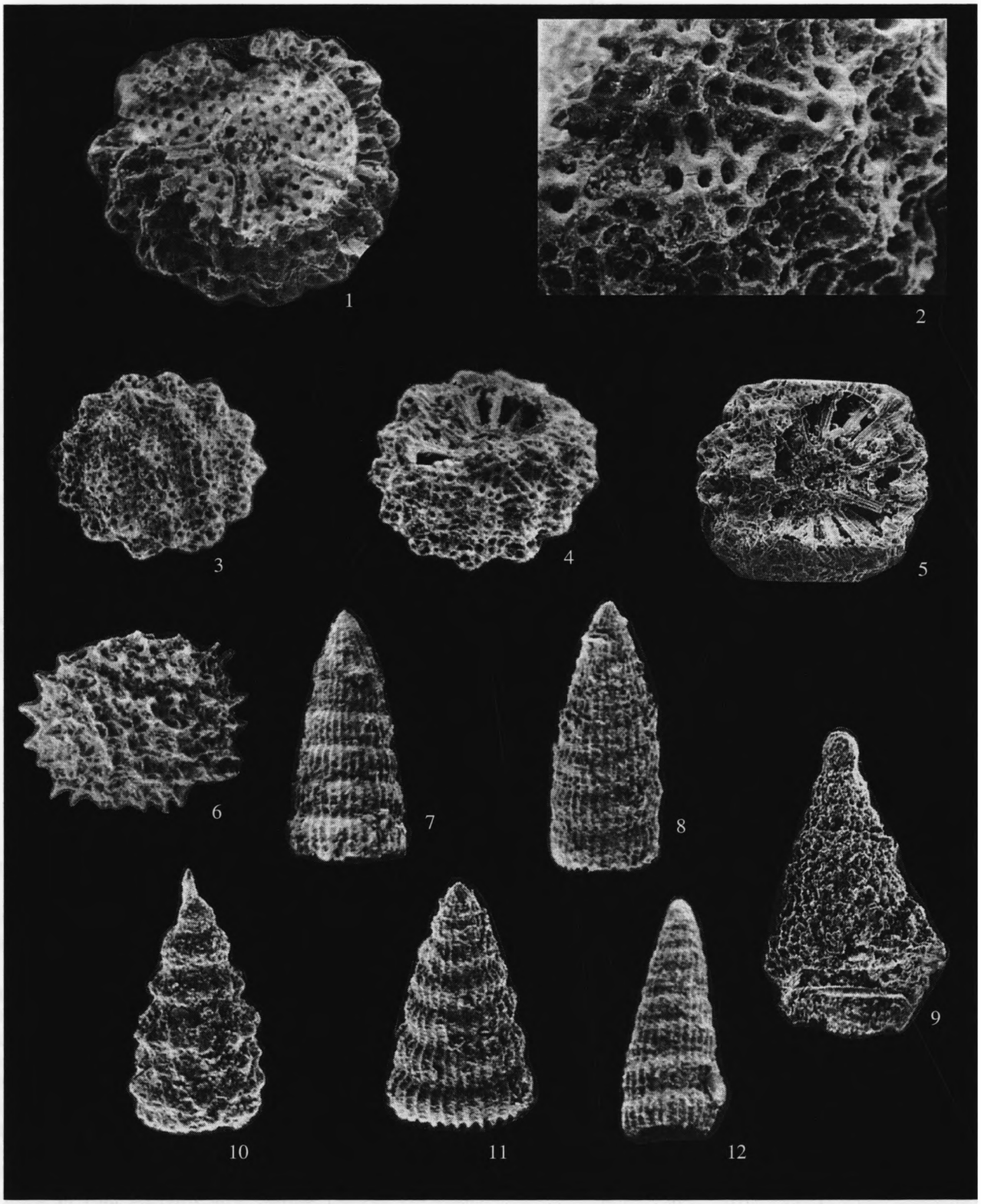


Plate III. (1–5) *Praeconocaryomma universa* Pessagno from samples 22, ×100 (1), 25, ×150 (3), 23, ×150 (4, 5), and details of perforations in test shown in fig. 4, ×600 (2); (6) *Praeconocaryomma lipmanae* Pessagno, Sample 22, ×150; (7, 8) *Archaeodictyomitra* sp. ex gr. *A squinaboli* Pessagno, Sample 22, ×150; (9) *Amphipyndax stocki* (Campbell et Clark), Sample 22, ×250; (10) *Xitus asymbatos* (Foreman), Sample 23, ×150; (11) *Dictyomitra multicostata* Zittel, Sample 22, ×150; (12) *Dictyomitra densicostata* Pessagno, Sample 22, ×120.

spines are 40–80 μm long, and pores range from 8 to 12 μm in diameter.

Comparison. In distinction from other species of the genus, arms of *Paronaella tumida* (Lipman, 1952) are swollen and rounded in plane. Shorter radial spines of specific shape differ the described species from *P. santonica* (Lipman).

Distribution: Coniacian–Campanian of the Russian plate and West Siberia.

Material: dozens of specimens.

Genus *Patulibracchium* Pessagno, 1971

Patulibracchium ingens (Lipman, 1952)

Plate II, fig. 4

Rhopalastrum ingens: Lipman, 1952, p. 37, Plate II, fig. 13 *Patulibracchium inaequalum*: Pessagno, 1971, p. 33, pi. 4, figs. 3-6; pi. 5, fig. 1

Holotype: no. 26/28, collection no. 6999, F.N. Chernyshev Central Geological Museum; Kuznetsk locality of Penza oblast, Santonian (Russian plate).

Description. Large tests have three arms, one of which can be sometimes longer than other arms. Central, slightly elevated region is approximately equal in size to the arm basal diameter. Arms are almost equally wide at proximal ends. In some specimens, wider sagittate terminations are characteristic of two longer arms, while in the others they characterize the shorter arm. In their widest part, sagittate terminations are swollen. Toward the distal end crowned sometimes by a short spine, arm walls converge. Brachiopyle is most distinct in the short arm. Test surface is perforated by round to subquadrangular pores, which can be arranged into rows oriented along the arm. At the junction point of several pores, skeleton can be decorated by small low hemispherical or conical tubercles. Pentagonal to hexagonal arrangement of pores is characteristic of the arm sagittate terminations.

Dimensions. Shorter and longer arms are170–200 and 600–500 μm long, respectively.

Comparison. In distinction from *P. teslaensis* Pessagno (Pessagno,1971), arms of the described species are oriented toward vertices of equilateral but not isosceles triangle.

Distribution: Cenomanian of California, Coniacian–Campanian of the Russian plate, upper Cenomanian *Triactoma parva–Patulibracchium ingens* Beds and lower Turonian and *Alievium superbum* Zone of the Crimea Mountains.

Material: 17 specimens.

ACKNOWLEDGMENTS

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Foraminifers and Depositional Environments of Upper Paleocene Sediments in the Northeastern Peri-Tethys

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Abstract—In the northeastern Peri-Tethys, the Goryachii Klyuch facies of the upper Paleocene are much more widespread than it was believed before. The facies correspond to dark colored, predominantly carbonate-free sediments bearing a specific foraminiferal assemblage with dominating agglutinated forms. These sediments accumulated in epicontinental basins, where the near-bottom water layer was likely contaminated by carbon dioxide that resulted in dissolution of calcareous foraminiferal tests.

Key words: foraminifers, Goryachii Klyuch facies, contamination by carbon dioxide.

Upper Paleocene deposits of the northwestern Caucasus are represented by dark colored carbonate-free shales and siltstones of the well-known Goryachii Klyuch Formation. Being known under various names in different structural–facies zones, similar deposits appear to be of the same type, as one can judge from their monotonous dark coloration, absence or very low content of carbonates, and from a specific composition of foraminiferal assemblage with dominating agglutinated forms characteristic of the rocks. In this work, I consider all deposits of this kind under the collective name of the Goryachii Klyuch facies.

Foraminiferal fauna of the Goryachii Klyuch Formation was first described by Glessner (1937) who discovered it in the Anapa section. Later on, after investigation of many sections situated between the Kuban River and town of Anapa, Subbotina (1950) published a summary characterization of fauna in question based on all the data available at that time.

In the northwestern Caucasus, I studied relevant sections exposed along the Mokryi Shibik, Ubin, Psekups, Belaya, Malyi Zelenchuk, Kuban, and Kurkuzhin rivers, and reexamined as well foraminifers collected by E.Yu. Zakrevskaya in the Nasypkoi Ravine of the eastern Crimea (Fig. 1). In western areas of the Ciscaucasia (Mokryi Shibik, Ubin, and Psekups sections), Paleocene deposits are of the flyschoid type. The Goryachii Klyuch Formation resting here on the lower Paleocene Tsitse Formation underlies the Il' Formation. In eastern areas, it is sandwiched between the El'burgan and Abaza formations. Thickness of the formation proper and its analogues is decreasing from western to eastern areas. The estimated thickness values range from 350 m (Mokryi Shibik section) and 430 m (Ubin River) to 100 m (Malyi Zelenchuk and Kuban rivers), 10 m (Belaya River), and even 3.5 m (Kurkuzhin site). The maximum thickness of 840 m is established in the Abin–Ubin interfluve (*Stratigraphy of the USSR...*, 1975). The exposures I studied have been already described in detail, like many other natural and borehole sections, in works by Grossgeim (I960), Shutskaya (1960b), Leonov and Alimarina (1964), and I omit their description in this work.

In the Kuban River section, rocks of the formation are still calcareous in the lowermost interval of few meters thick and yield planktonic foraminifers of the *Morozovella conicotruncata* Zone of the Crimea–Caucasus scale. In the middle interval of the unit, Shutskaya (1970) distinguished the *Igorina tadjikistanensis djanensis* Zone, whereas in the upper part of many sections there is a thin bed of marl (usually less than 1 m thick) that yields planktonic foraminifers, which are not diverse and typical of the *Acarinina subsphaerica–A. acarinata (=A. nitida)* Zone. According to personal communication of N.G. Muzylev, the boundary separating the nannoplankton zones NP 8 (*Heliolithus riedelii*) and NP 9 (*Discoaster multiradiatus*) is detected exactly in this bed of the Kuban River section. Above this bed there is a sapropelic layer a few centimeters thick, which bears abundant planktonic foraminifers of the *Acarinina acarinata (=A. nitida)* Zone and has been recently regarded as boundary horizon between the Paleocene and Eocene. Accordingly, the Goryachii Klyuch facies span at maximum the stratigraphic interval from *Morozovella conicotruncata* to *Acarinina acarinata (=A. nitida)* Zone.

Fossils in the Goryachii Klyuch facies are represented by foraminifers, nannoplankton (in carbonate interlayers), radiolarians, mollusks (rare specimens), and by dinocysts. The assemblage of foraminifers studied better than others consists predominantly of agglutinated species, mostly of morphologically simple

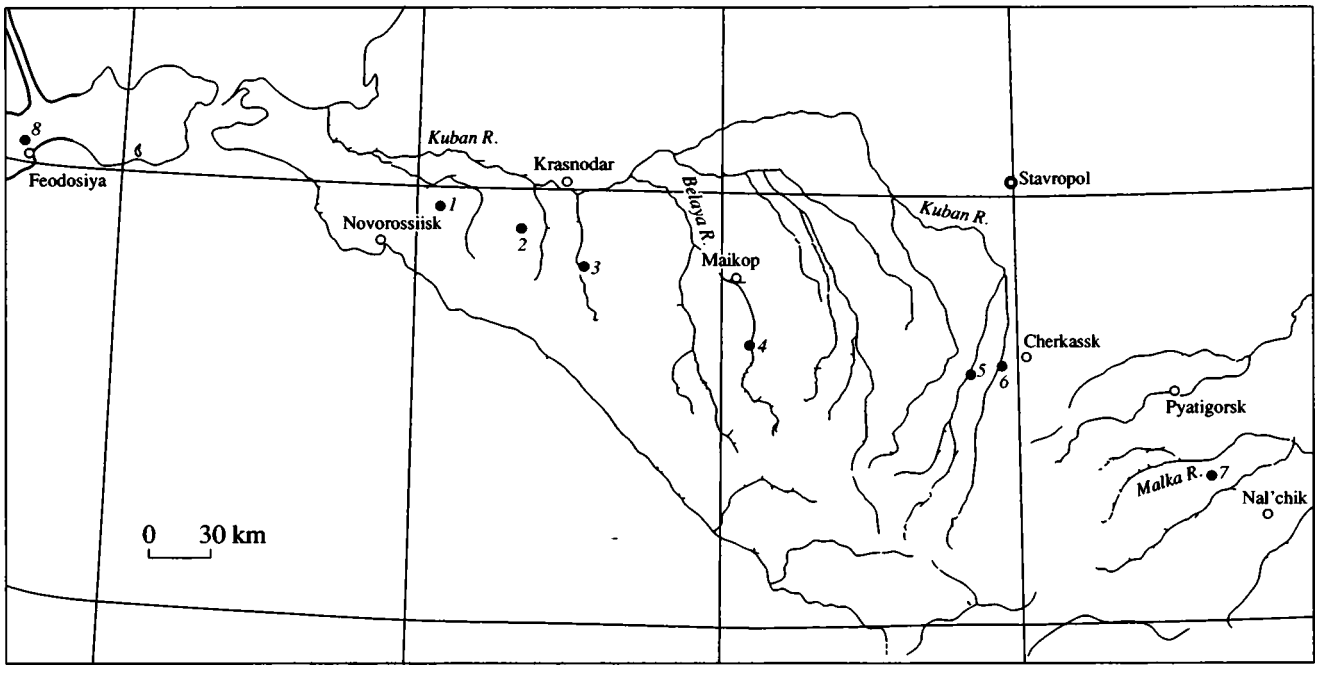

Fig. 1. Localities of studied sections: (*1*) Mokryi Shibik R.; (*2*) Ubin R.; (*3*) Psekups R.; (*4*) Belaya R.; (*5*) Malyi Zelenchuk R.; (*6*) Kuban R.; (*7*) Kurkuzhin R.; (*8*) Nasypkoi Ravine.

forms from the orders Astrorhizida and Ammodiscida (Figs. 2–8).[1]

Like previous researchers, I failed to discriminate biostratigraphic subdivisions based on benthic foraminifers. In general, the foraminifers can be regarded as a uniform assemblage, composition of which did not change practically during the late Paleocene.

In contrast to the common opinion that the Goryachii Klyuch Formation and its analogues bear almost exclusively the agglutinated foraminifers, my analysis of encountered assemblages showed that planktonic and calcareous benthic forms coexists in all the sections, and that the secretory benthic foraminifers are sufficiently diverse in taxonomic aspect. As for agglutinated forms, they prevail over the others owing to a greater number of individual species but not because of their greater abundance rate. Subbotina (1936) who came to the same conclusion long ago wrote that being enriched in agglutinated taxa the foraminiferal assemblage of the Goryachii Klyuch Formation is more diverse that characterizing the underlying marls of the El'burgan Formation.

It is remarkable that the western Mokryi Shibik section situated in the flysch zone of the northwestern Caucasus is completely depleted in calcareous material. Nevertheless, agglutinated foraminifers are dominants even here.

At the top of the Goryachii Klyuch Formation and its analogues, there is a marl bed frequently separated by an erosion surface and corresponding to the Laba Formation base in the scheme of Leonov and Alimarina (1964). Characteristic of the bed are the diversity of dominant calcareous benthos and the more frequent occurrence of planktonic forms. Within the northwestern Caucasus, this level is correlative with slightly calcareous varieties of the Goryachii Klyuch facies in eastern areas, where E.A. Shcherbinina (personal communication) who studied exposures along the Kurkuzhin River identified nannoplankton of NP 8 Zone. Unfortunately, zonal assemblage of planktonic foraminifers has not been encountered here. The bed in question is overlain by a sapropelic bed with dominant planktonic foraminifers.

Two subdivisions overlying the Goryachii Klyuch Formation and its analogues are the Il' Formation on the west (shales with sandstone interlayers and chalk inclusions) and the Abaza Formation on the east (the same shales with cherty interlayers). Both formations are free of carbonate material, yield a uniform assemblage dominated by agglutinated forms, and can be also attributed to the Goryachii Klyuch facies. Deposits overlying them are calcareous, bearing fauna of dominant calcareous benthos characteristic of shelf seas.

Radiolarians represent the other group of microfauna typical of the Goryachii Klyuch facies. These

[1] In the Psekups River section, where only basal beds of the formation have been studied, they yield *Ammodiscus tenuissimus* Grzyb., *Hormosinella? velascoensis* (Cush.), *Glomospira irregularis* (Grzyb.), *Repmanina charoides* (Jones et Parker), *Textularia agglutinans* Orb., *Lituotuba lituiformis* (Brady), *Spiroplectammina spectabilis* (Grzyb.), *Clavulina parisiensis* Orb., *Haplophragmoides subsphaeroides* Subb., *H. tenuis* Cush., and *Subbotina triloculinoides* (Plumm.)

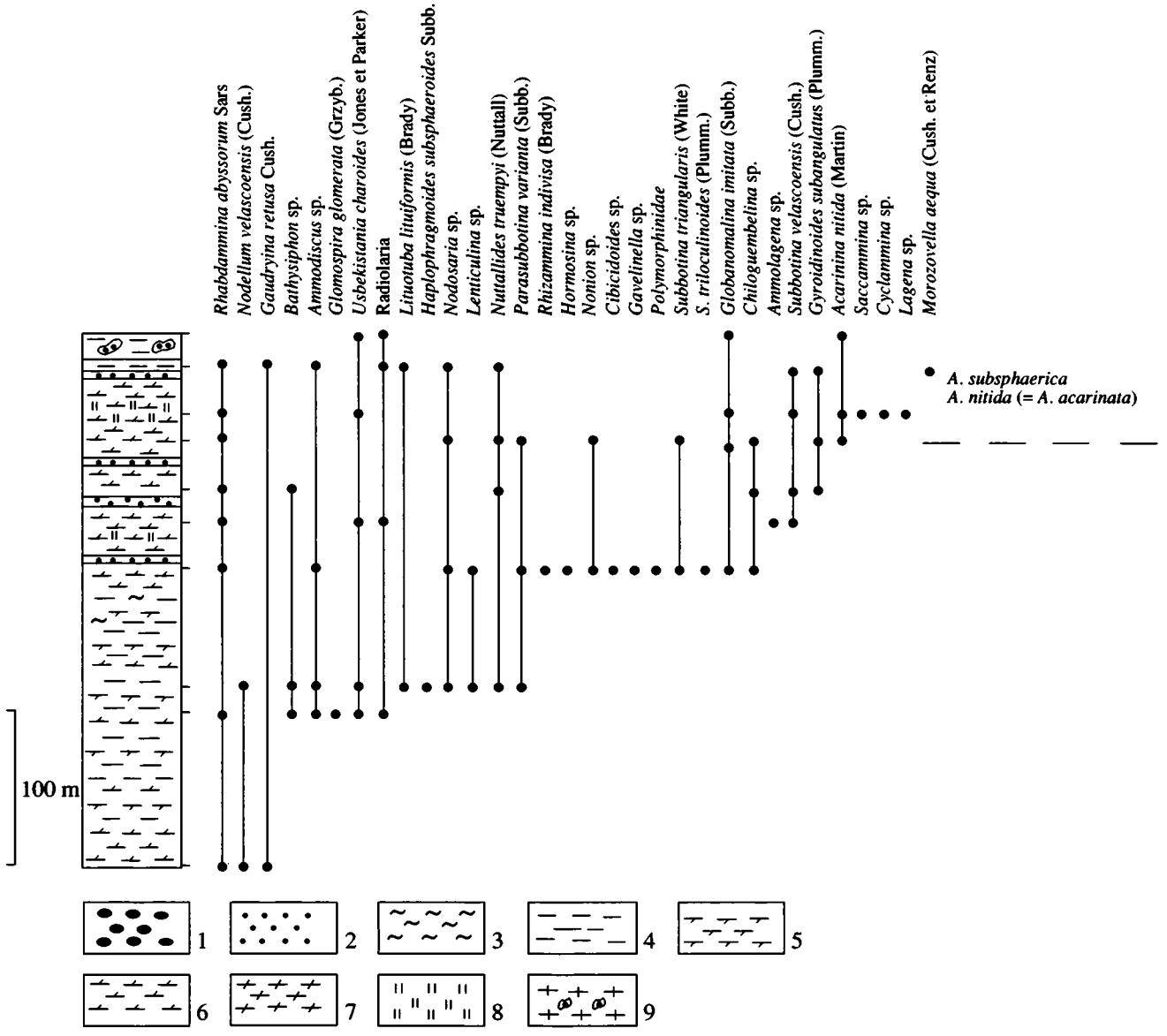

Fig. 2. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, Mokryi Shibik section: (1) gravelstone; (2) sandstone; (3) siltstone; (4) carbonate-free shale; (5) slightly calcareous clay; (6) highly calcareous clay; (7) marl; (8) opoka; (9) nummulitic limestone.

fossils occur almost everywhere in the study region being more abundant than planktonic foraminifers. Radiolarian tests usually are highly recrystallized, and in my own collection of samples, they are practically undeterminable.

It was believed formerly that the Goryachii Klyuch facies are present only in western areas of the Ciscaucasia. Nevertheless, published data imply a wider geographic range of these sediments (Fig. 9). The facies are known in the Black Sea region (Konenkova, 1979), in the Boltysh depression of the Ukrainian Shield southern flank, which is regarded as an astrobleme of Cretaceous age (Boĭko *et al.*, 1985), and in the Dnieper–Donets depression (Vasilenko, 1950; Kaptarenko-Chernousova, 1960; Zhmur *et al.*, 1969). Sediments of this kind have been described as well in the southeastern Crimea (Shutskaya, 1960a; Bugrova *et al.*, 2002; and my data, see Fig. 8), in the Kerchenskii Peninsula (Voloshina and Denega, 1969), Scythian plate (Nikitina and Shvemberger, 1963), northwestern Circum-Caspian depression (Shutskaya, 1960), and in eastern Azerbaijan (Khalilov, 1948, 1959). In southern Russia, the eastern limitation of facies distribution area approximately corresponds to the Kurkuzhin–Artezian (Kuma River mouth) line (Shutskaya, 1960b) that demarcates the Malka ledge, which likely acted as a hydrologic and geochemical barrier in the past. In Asia, the Goryachii Klyuch facies occur in central and southern segments

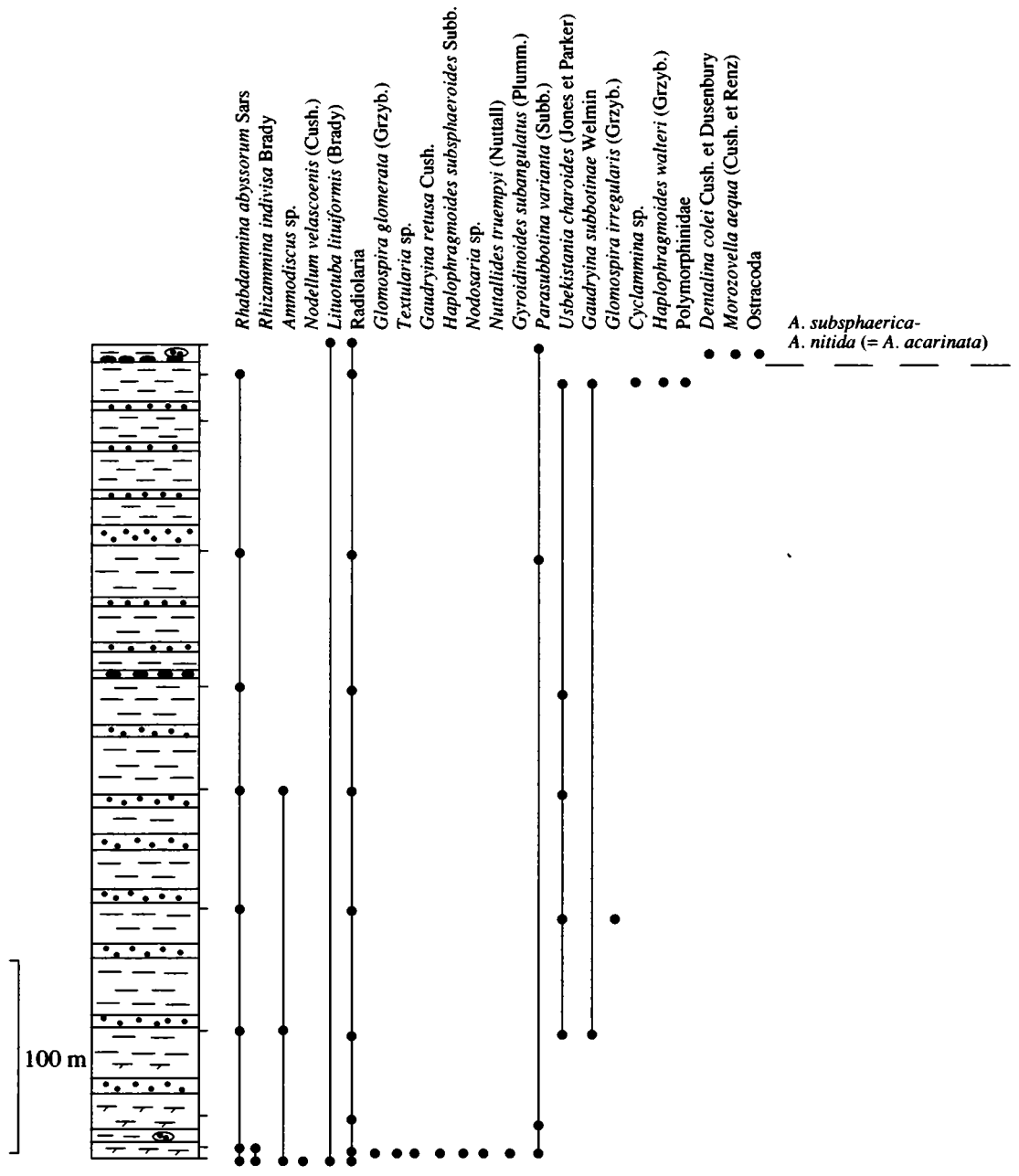

Fig. 3. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, Ubin River section (symbols as in Fig. 2).

of the Transurals (Vasil'eva and Malyshkina, 2003), in the Buzachi Peninsula (Bykova, 1960), Malyi Balkhan (Khalilov, 1948), western Kopetdag (Morozova *et al.*, 1967), Kyzylkum (Tsatsir, 1970), and in the Tajik (Bykova, 1953a) and Fergana (Bykova, 1953b) depressions. In addition, the upper Paleocene deposits of Egypt enclose interlayers of black carbonate-free shales with agglutinated foraminifers, which are very similar to their counterparts from the Goryachii Klyuch Formation (Krasheninnikov, 1969). Unfortunately, the initial time of the facies accumulation is recorded in a few areas, because zonal assemblages of calcareous plankton are absent in the others. For instance, the formation base corresponds to the *Morozovella conicotruncata* Zone in the Kuban River section. In the Black Sea region on the other hand, carbonate deposits of the Mirnoe Formation correlative with the marl bed of the Kuban River area are overlain by dark colored sediments of the Ochakovo Formation (Goryachii Klyuch facies), which belong to the *Acarinina acarinata* (= *A. nitida)* Zone (*Unified...*, 1987). Accordingly, the incipient accumulation time of the Goryachii Klyuch facies was asynchronous in different parts of the paleobasin during the late Paleocene. In contrast, termina-

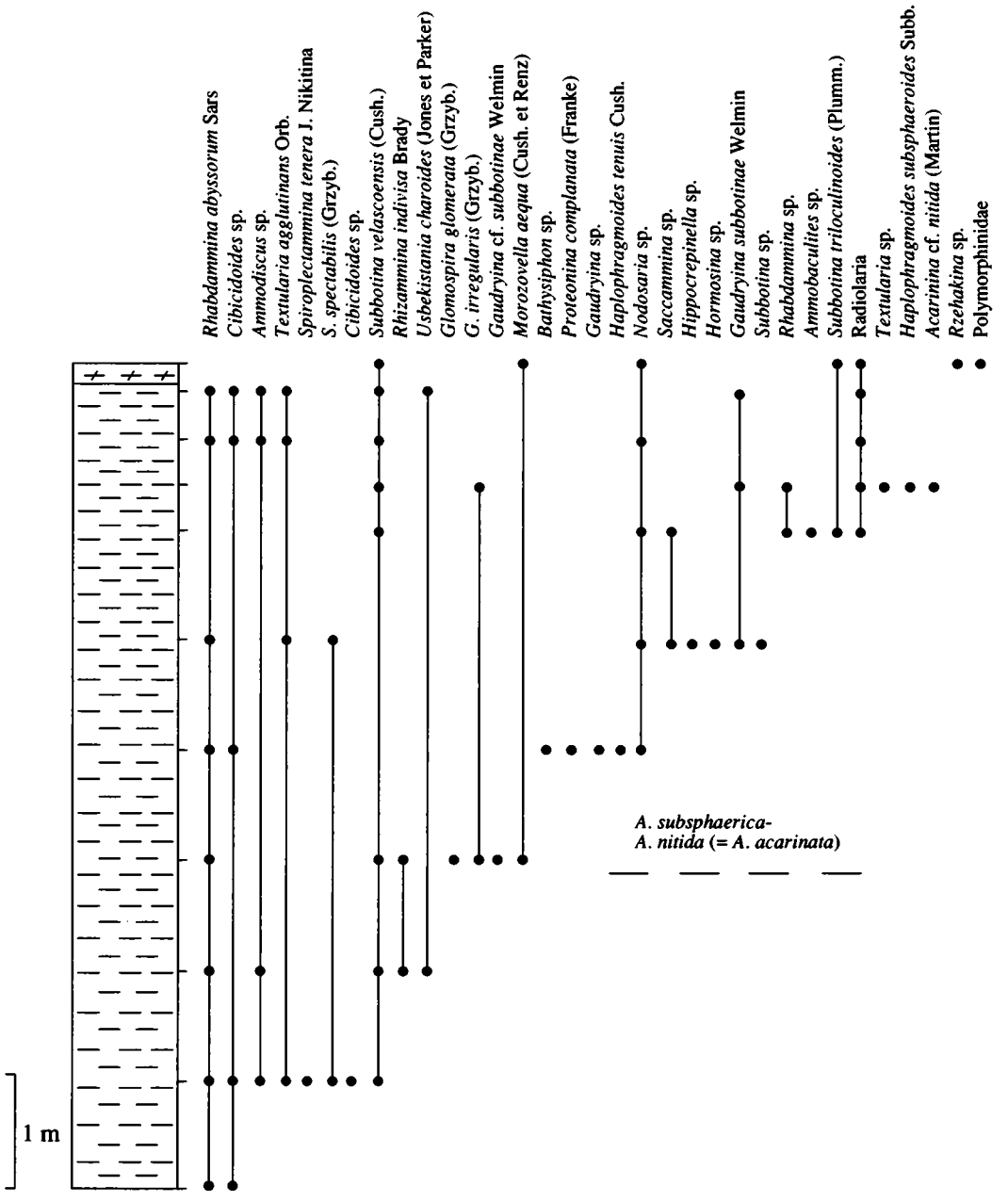

Fig. 4. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, Belaya River section (symbols as in Fig. 2).

tion of the facies deposition coincides with a hiatus that is recorded in all southern sections of the former USSR (Muzylev, 1996).

As for depositional environments of the Goryachii Klyuch facies, they have been interpreted variously.

Subbotina (1950) who regarded the deep-water origin of the facies as hardly probable arrived at the conclusion, based on dwelling conditions of present-day agglutinated foraminifers, that respective areas of the paleobasin were under influence of cold currents. Her conclusion is consistent with observation, according to which radiolarians replace planktonic foraminifers in contemporary cold seas. Bykova (1953b) who studied sections of the Fergana depression suggested that development of calcareous benthos was suppressed in the past cold-water basin, because abundant organic matter, the decomposition product of sea vegetation, emanated carbon dioxide creating conditions favorable for dissolution of $CaCO_3$. Grossgeim (1960) assumed a weak reducing state of waters in the paleobasin of the western Ciscaucasia. Beniamovskii, Shcherba, and Kopp (Beniamovskiĭ and Shcherba, 1995; Kopp and Shcherba, 1998) argued for a considerable basin depth comparing the late Paleocene foraminifers with

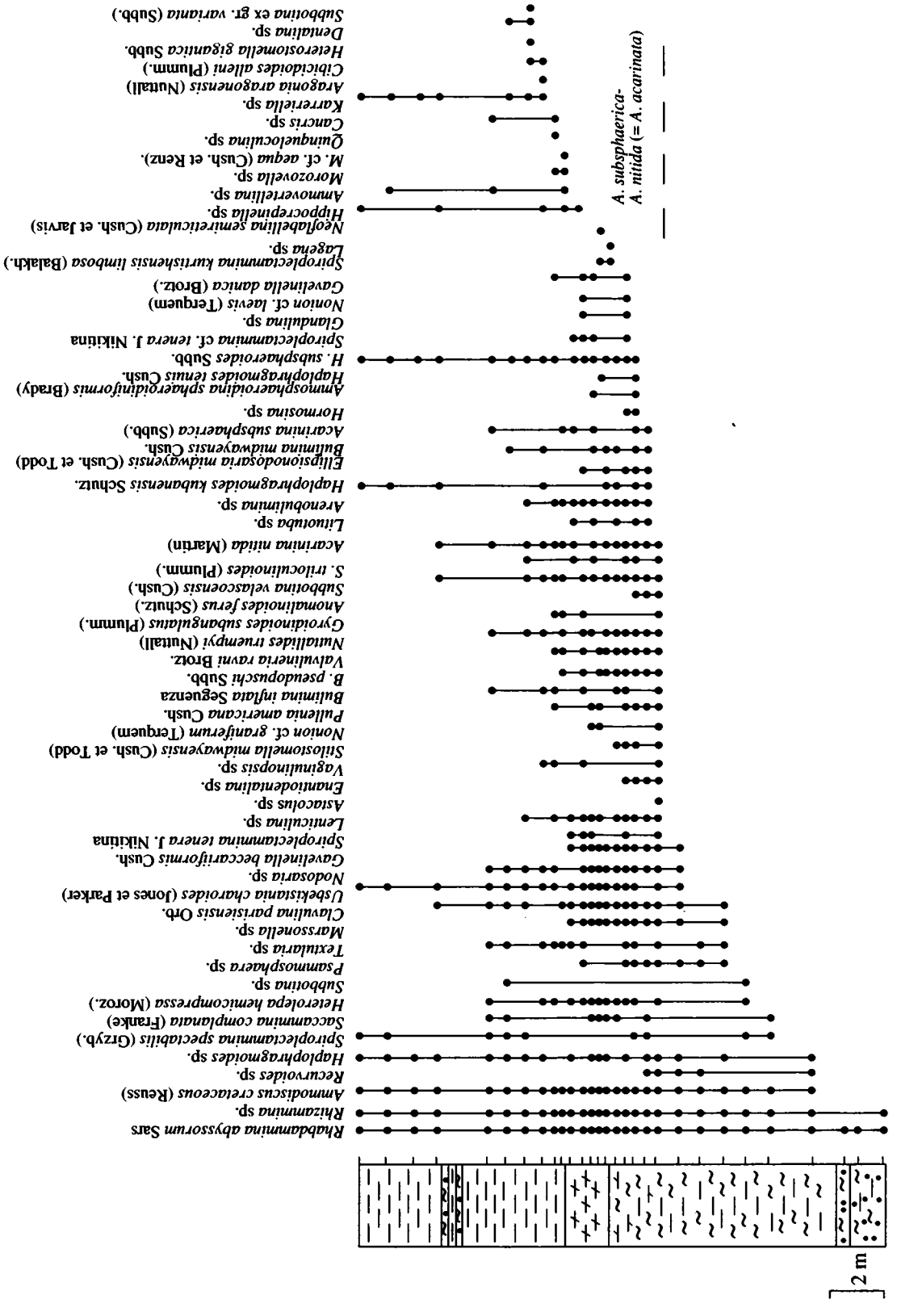

Fig. 5. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, Malyi Zelenchuk section (symbols as in Fig. 2).

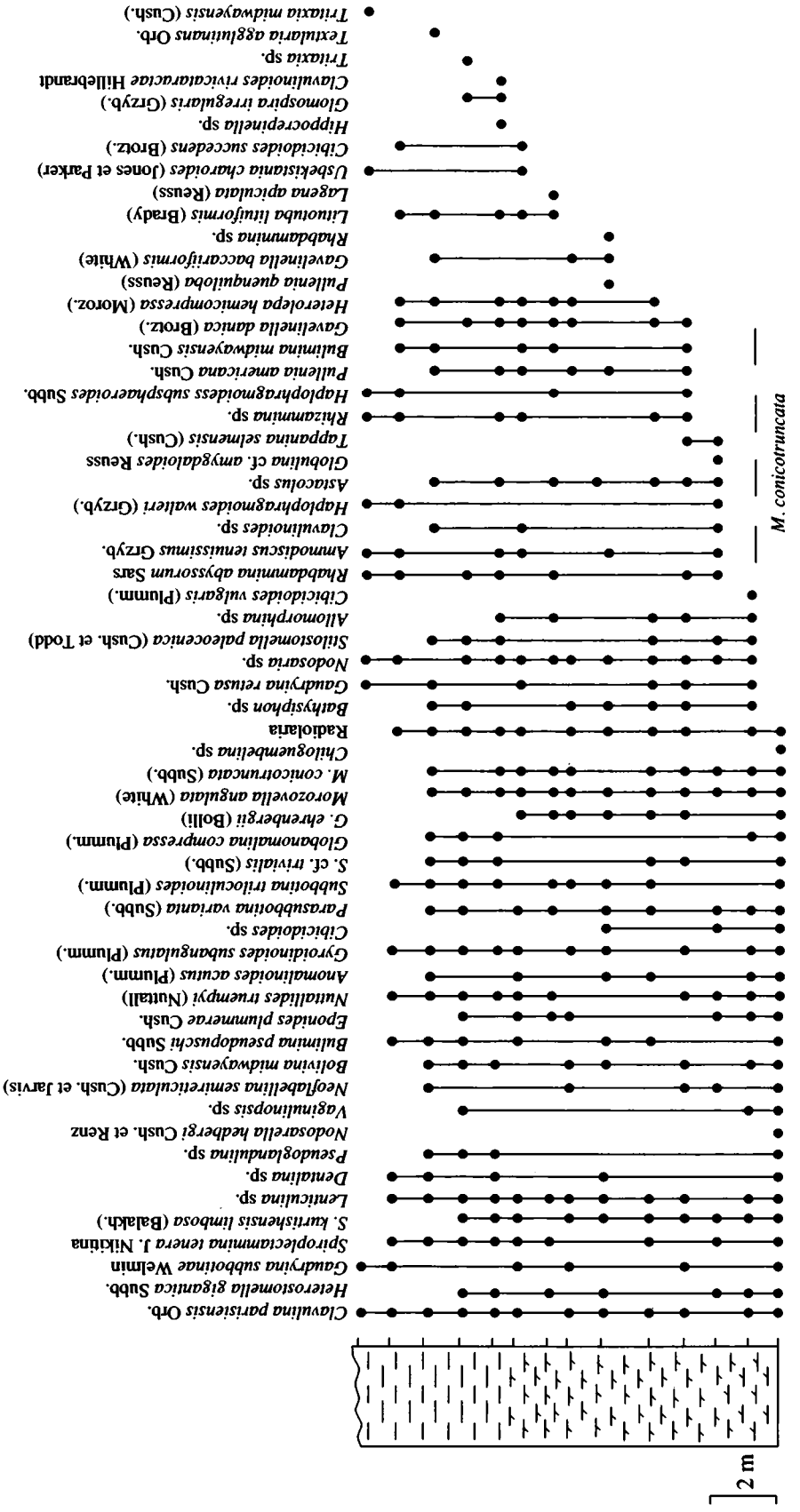

Fig. 6. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, Kuban River section (symbols as in Fig. 2).

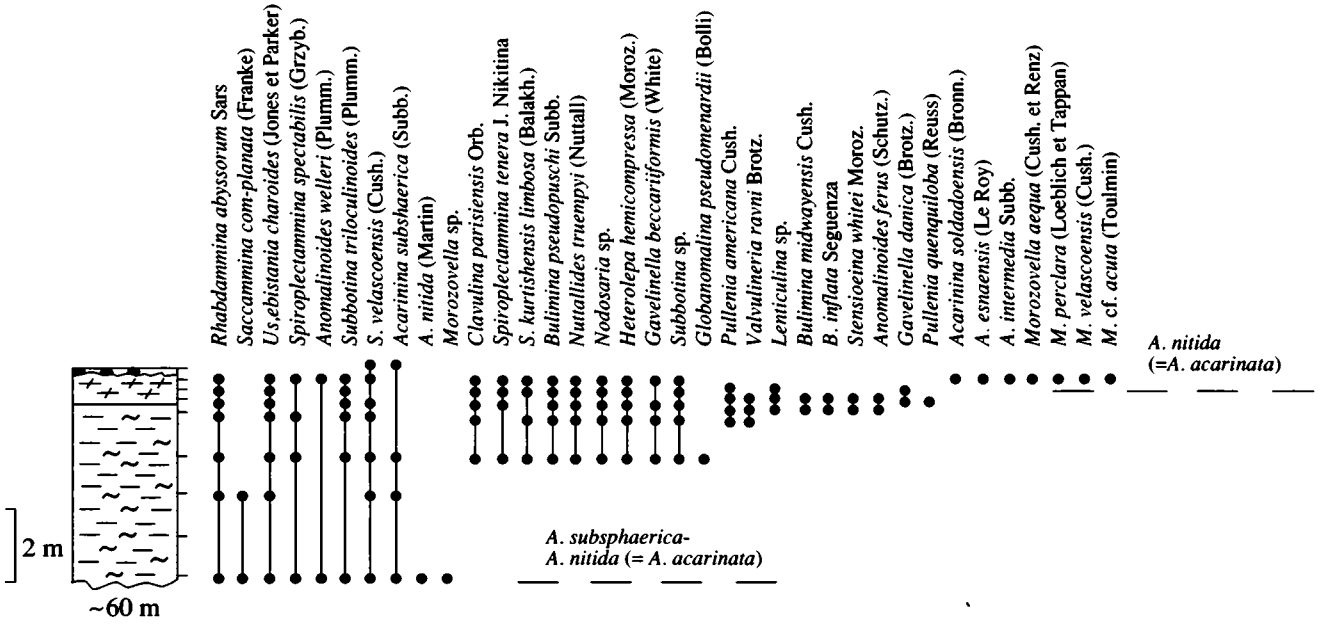

Fig. 6. (Contd.)

present-day foraminiferal faunas of deep-water settings.

Recently Beniamovskii and Shcherba (1999) proposed the model, according to which a deep-water trough formed in the western part of the Greater Caucasus at the commencement of the late Paleocene. Characteristic of the continental slope and rise of the trough were the flysch deposition, stagnation of waters, and "deep-water anoxic" coenoses (? S.S.) of foraminifers. The middle Thanetian transgression triggered the upwelling in that trough adjacent to the epicontinental basin and transport of water mass, which, being depleted in oxygen, ascended from the epibathyal zone to the shelf and carried the "anoxic" assemblages of agglutinated foraminifers.

Analyzing the available materials, I feel it necessary to specify the formation environments of the Goryachii Klyuch facies. As for the water depth in paleobasin, it was naturally not uniform in different areas, probably attaining 300–350 m in the deep shelf zone. The marl and sapropelic beds present in the Goryachii Klyuch Formation are of transgressive character in places (Leonov and Alimarina, 1964) and yield the shallow-water nannoplankton (Muzylev, 1994) and benthic foraminifers characteristic of the neritic zone. In general, many foraminifers from the facies under consideration are represented by species typical of the Midway shelf fauna. These are *Tritaxia midwayensis* (Cush.), *Bulimina midwayensis* Cush. et Parker, *Nonion graniferum* (Terquem), *Gyroidinoides subangulatus* (Plumm.), *Cibicidoides succedens* (Brotzen), *C. alleni* (Plumm.), *Gavelinella danica* (Brotzen), *Anomalinoides acutus* (Plumm.), *Stilostomella paleocenica* (Cush. et Todd) and others. This fauna was originally described from the Midway Formation of Texas (Plummer, 1927) and then it was traced worldwide by Berggren and Aubert (1975). Nevertheless, it includes *Gavelinella beccariiformis* (White) and *Nuttallides truempyi* (Nuttall), the species known in the Velasco fauna attributed to coenosis of a continental slope (type locality is the Velasco Formation of Mexico). Thus, it is unlikely that the basin was deeper than the carbonate compensation depth, and this conclusion is consistent with a considerable proportion of calcareous foraminifers (about 40%) in the Mokryi Shibik section situated in the flysch zone, where its strata are slightly calcareous though not all of them. The phosphorite horizon known in central area of the western Ciscaucasia (Leonov and Alimarina, 1964) is an additional argument in favor of above inference. The flyschoid character of sediments likely reflects in this case a dissected shelf relief rather than the continental slope and rise settings.

The peculiar assemblage of benthic foraminifers is, in my opinion, a consequence of bottom water contamination by carbon dioxide. According to geochemical data on the lower, still calcareous part of the Kuban River section (personal communication of Yu.O. Gavrilov), C_{org} content in sediments ranges from 0.1 and less to 1.7%, being as great in average as 0.3%. Taking into consideration that sapropelic sediments contain 0.5 to 2% of C_{org} by definition of participants of "Glomar Challenger" Leg 42 (Kidd *et al.*, 1978), I think that carbon dioxide emanation from abundant organic matter (its source is problematic so far) led to dissolution of calcareous tests under conditions of weakly aerated near-bottom water. The dissolution effect is directly evident from corrosion marks frequently observable on tests of secretory benthic forms. The indirect evidence

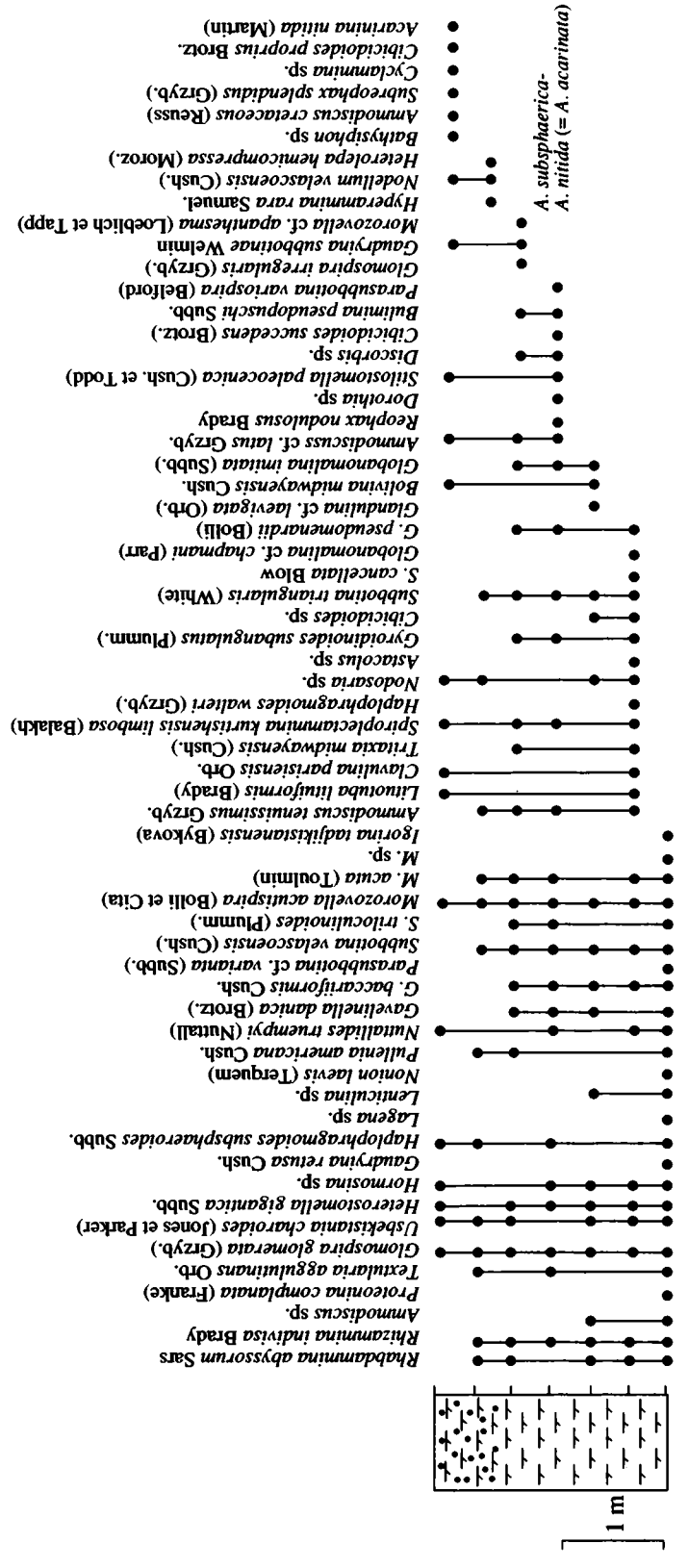

Fig. 7. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, Kurkuzhin River section (symbols as in Fig. 2).

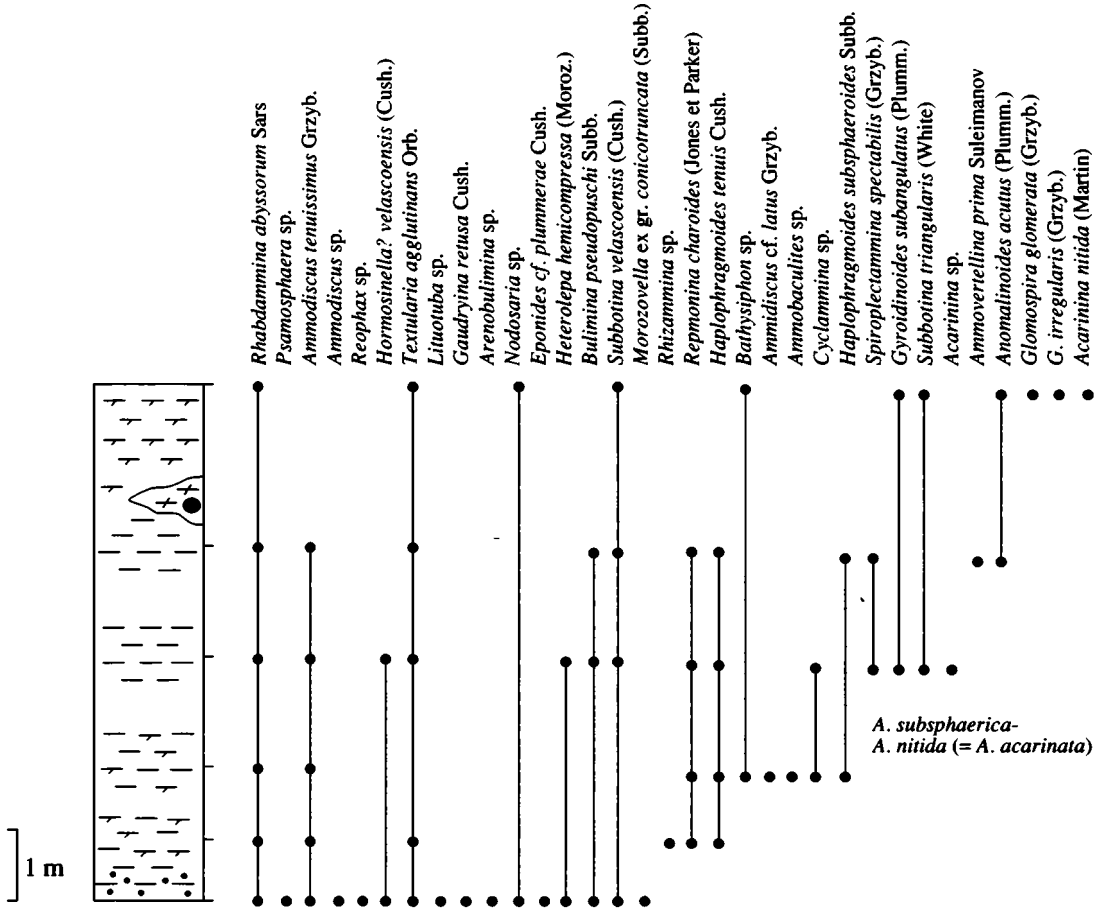

Fig. 8. Lithology and stratigraphic ranges of foraminifers in the Goryachii Klyuch facies, section of the Nasypkoi Ravine (symbols as in Fig. 2).

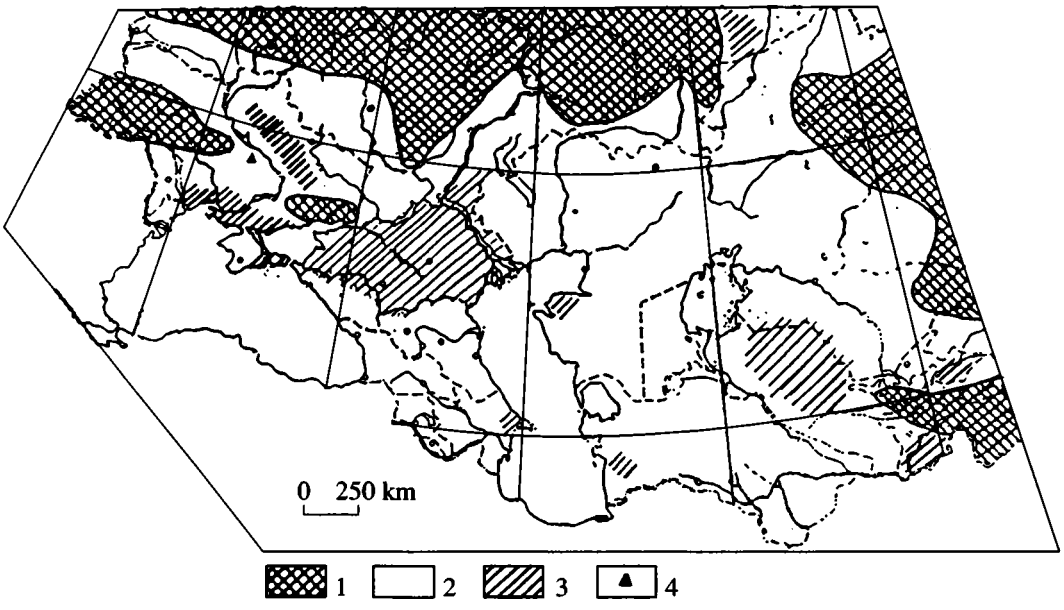

Fig. 9. Late Paleocene paleogeography of the northeastern Peri-Tethys: (1) land; (2) sea; (3) known distribution areas of the Goryachii Klyuch facies; (4) Boltysh depression.

is presence of Fe- and Mg-carbonate concretions and of disseminated Ca-carbonate in the studied sediments (Bykova, 1953b; Grossgeim, 1958; Voloshina and Denega, 1969). Similar peculiarities are characteristic, for instance, of recent sediments in the Barents Sea (Digas, 1980; Khusid, 1989; Korsun *et al.*, 1994) and near the southwestern coast of South America (Khusid, 1984).

Abundance of concurrent radiolarians (Subbotina, 1950; Borisenko, 1960) was possibly not a consequence of outburst in their development, but a result of low dissolution rate of silica in the weakly acid medium.

Perturbations in the near-bottom water layer should be sufficiently active for influx of oxygen, which is a component of carbonic acid, and weak enough to keep the latter after its formation near the sea floor. Such a balance was interrupted by transgression that resulted in deposition of the marl bed. After the subsequent regression and erosion of sediments, the process resumed and stimulated deposition of sapropelic sediments (for depositional environments of the latter, see Stupin and Muzylev, 2001, and references therein). The next period of gradual stabilization was followed by the new intense regression and erosion. The last event recorded in the sections corresponds to the early Ypresian transgression.

CONCLUSION

(1) During the late Paleocene, the northeastern Peri-Tethys represented a system of epicontinental basins, which were not deeper than 300–350 m.

(2) The Goryachii Klyuch facies occupying large southern areas of the former USSR are of a much wider geographic range than it was believed formerly. The incipient stage of their deposition that lasted through the late Paleocene was not synchronous in different basin areas, and the terminal stage corresponds to the regional hiatus.

(3) Agglutinated forms are dominants of foraminiferal assemblage from the Goryachii Klyuch facies not because of their diversity but owing to the greater abundance rate of individual species. They persistently associate with calcareous benthos that is of a quite diverse taxonomic composition.

(4) The peculiar paleocoenosis of foraminifers was likely controlled by carbon dioxide contamination of near-bottom water layer, and calcareous tests experienced posthumous dissolution under these conditions.

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Correlation of the Pliocene–Quaternary Foraminiferal and Nannofossil Zonations in the North Atlantic

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Abstract—A thorough investigation of calcareous plankton from the Pliocene–Quaternary sediments of DSDP Sites 397, 608, 609, and 611 enabled correlation between the foraminiferal and nannofossil zonations. The first and last occurrence levels of nannoplankton and foraminifer index species are defined and calibrated against the chronological scale by Berggren *et al.* (1995). First occurrence of *Globorotalia margaritae* is recorded in studied sections at 6.1–6.3 Ma and last occurrence of *Discoaster quinqueramus* corresponds to the level of 5.94 Ma. The first appearance of *Gephyrocapsa oceanica* is established at 1.87 Ma. The distinct correlation between the planktonic foraminiferal and nannofossil zonations is revealed in the North Atlantic region between 26° and 52°N.

Key words: Pliocene, Quaternary, foraminifers, nannofossils, correlation, zonations, datums, Atlantic.

INTRODUCTION

A central problem of stratigraphy, i.e., the detailed subdivision of sedimentary succession and the precise dating of stratigraphic units, can be solved by means of correlation of different biostratigraphic zonations, for instance, of zonations established for well-studied planktonic foraminifers and nannoplankton. The Pliocene–Quaternary zonations of these two groups of planktonic calcareous fossils are of a high resolution, tested in different regions of the World Ocean. As the boundaries between discriminated biostratigraphic units mostly do not coincide, the parallel analysis of two or more fossil groups leads to an enhanced stratigraphic resolution and to the more exact dating of strata.

The nannofossil and planktonic foraminiferal zonations have been correlated in some previous works (Krasheninnikov, 1978, 1980; Berggren *et al.*, 1985, 1995; Krasheninnikov *et al.*, 1999) though without proper attention to critical datum levels of microfauna and flora established in the same borehole sections and samples. In the above-mentioned works, the top of the nannoplankton *Discoaster quinqueramus* Zone NN11 (CN9) is placed below the first occurrence of *Globorotalia margaritae* or at the same level. The base of the *Discoaster brouweri* Zone CN12 (NN16–18) is established either within the *Globorotalia margaritae* Zone or at the top of this zone. Moreover, the detailed correlation with due regard for subzones discriminated by Okada and Bukry (1980) and for the high-resolution Quaternary nannofossil zonation (Gartner, 1977) has not been considered.

In this paper we discuss the correlation of the Pliocene–Quaternary sediments penetrated at DSDP Sites 397, 608, 609, and 611. All these sections have been studied by paleomagnetic methods. In her study of nannofossils, Golovina coordinated the results with the standard scale by Martini (1971), with the low-latitude zonation suggested by Okada and Bukry (1980), and with the Quaternary standard developed by Gartner (1977). Bylinskaya who studied planktonic foraminifers used for correlation the stratigraphic scale by Bolli and Saunders (1985) that was slightly modified not long ago (Bylinskaya *et al.*, 2002). The first and last occurrence levels (FO and LO, respectively) of most index species established in the studied sections are dated based on the chronological scale published by Berggren *et al.* (1995). The reported ages characterize events distinctly traceable in the sections and the levels of certain paleomagnetic reversals near the particular events.

STUDIED SITES AND STRATIGRAPHY

Southernmost DSDP Site 397 (26°50.7′ N, 15°10.8′ W, 2900 m) is located in the subtropical zone, on the continental slope of the northwestern African coast (Fig. 1). The Pliocene and Quaternary sediments accumulated here under combined hydrodynamic influence of the cold Canary Current and upwelling. The Pliocene sediments are 280 m thick and the Quaternary strata are about 140 m thick. Site 608 (42°50′ N, 23°05′ W, 3526 m) was drilled on the southern flank of the King's Trough complex, east of the Mid-Atlantic Ridge. The total thickness of the Pliocene and Quaternary deposits penetrated at the site is about 160 m. Site 609 (49°53′ N, 24°14′ W, 3883 m) is located in the eastern flank of the Mid-Atlantic Ridge, where a 350-m-thick sequence of

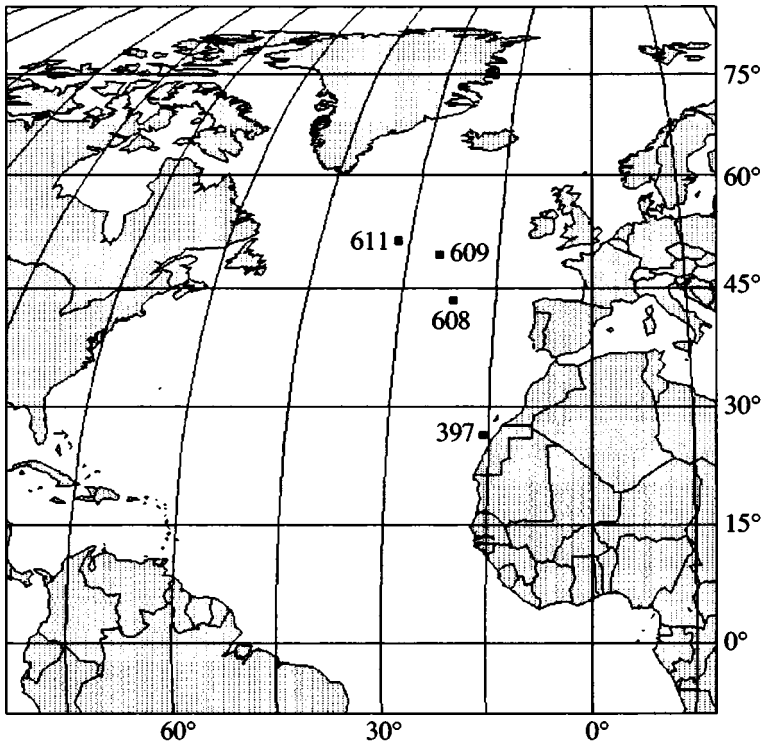

Fig. 1. Localities of the studied DSDP sites.

the Pliocene–Quaternary sediments was recovered. Site 611 (52°50′ N, 30°18′ W, 3230 m) is in a lower part of the Gardar Ridge southeastern flank, where sediments accumulated under influence of the deep-water overflow from the Norwegian Sea. The Pliocene and Quaternary deposits recovered here are about 330 m thick.

The Pliocene–Quaternary intervals of all the holes yield the diverse assemblages of planktonic foraminifers and nannofossils, which are suitable for a high-resolution biostratigraphic subdivision. The initial study of sediments during DSDP Legs 47 and 94 (von Rad *et al.*, 1979; Ruddiman *et al.*, 1987) was subsequently followed by the more thorough investigation (Krasheninnikov and Bylinskaya, 1999; Bylinskaya *et al.*, 2002). As the planktonic foraminiferal and nannofossil assemblages and zonal units recognized at the sites are described in detail in the last work, both issues are considered here in brief. It should be noted that zonal units are distinguished in this study based on their characteristic assemblage but not on individual key species. At the same time, the events of appearance and extinction of some forms in different climatic zones can serve as additional markers enabling a higher resolution of stratigraphic subdivision. The first and last occurrence datums defined for foraminiferal and nannofossil species most important in terms of stratigraphy are presented in the tables.

In our previous work (Bylinskaya *et al.*, 2002), we analyzed the calcareous plankton zonations in the Atlantic Ocean and proved the possibility to use the low-latitude zonations in higher latitudes, up to the boreal realm, although their resolution is different in various latitudinal belts. Analyzing zonation of planktonic foraminifers, we used the scheme by Bolli and Saunders (1985) that was detailed in the middle Pliocene interval.

The *Globorotalia margaritae, Globorotalia miocenica, Globorotalia tosaensis*, and *Globorotalia truncatulinoides* zones with subzones are distinguished in the Pliocene–Quaternary foraminiferal zonation at DSDP Sites 397, 608, 609, and 611. The nannofossil Zones CN10–CN12 with the corresponding subzones and five Quaternary biostratigraphic units by Gartner (1977) were recognized in addition.

RESULTS AND DISCUSSION

Site 397

Since the site is located in a warm-water region of the Atlantic, the nannofossil zonation was correlated herewith that of Okada and Bukry, although we failed to distinguish certain subzones, the index species of which are either lacking or show their earlier (later) occurrence in the section.

The Pliocene sediments are recovered by Cores 45 to 15 (Fig. 2). The base of the early Pliocene foraminiferal *Globorotalia margaritae* Zone is placed in the lowermost Core 45 (Bylinskaya *et al.*, 2002). The index species LO defines the zone upper boundary, being

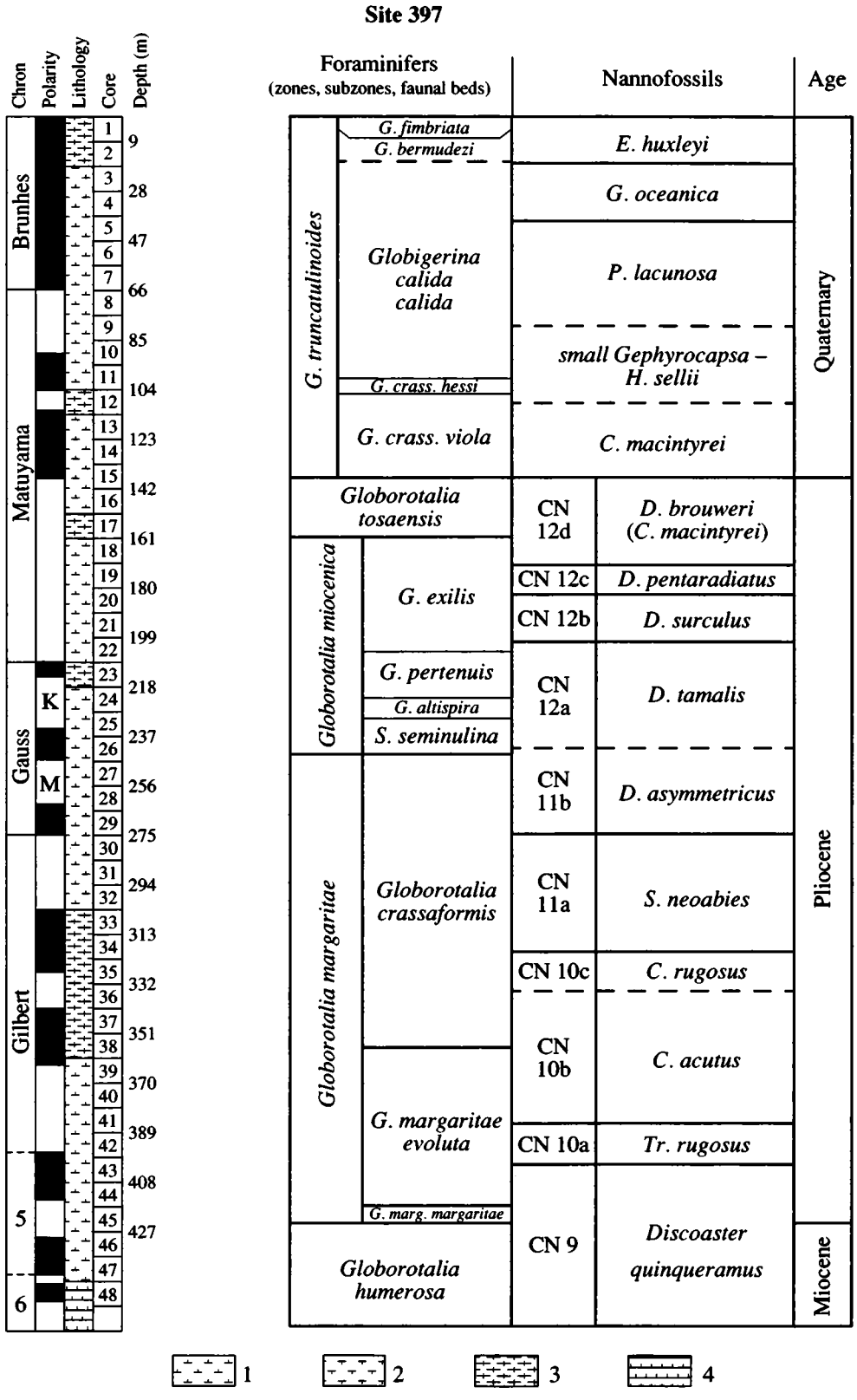

Fig. 2. Correlation between of the Pliocene–Quaternary zonal units of calcareous plankton from DSDP Site 397: (1) nannofossil, (2) foraminiferal, and (3) nanno-foraminiferal oozes; (4) nannofossil chalk; (K) Kaena Subchron; (M) Mammoth Subchron.

Table 1. First (FO) and last occurrence (LO) datums of planktonic foraminiferal species

Datum events	Age (Ma)			
	Site 397	Site 608	Site 609	Site 611
FO *G. calida calida*	1.05	1.07	0.99	
LO *G. crassaformis viola*		1.45	1.48	
FO *G. truncatulinoides*	1.95		1.95	
LO *G. extremus*	2.08		2.05	
LO *G. exilis*	2.30			
LO *G. miocenica*	2.40			
LO *Gq. altispira*	3.03	3.03		
FO *Sph. dehiscens*	3.13			
LO *Sph. seminulina*	3.14	3.10		
LO *G. margaritae*	3.23	3.47	3.92	
LO *Gq. dehiscens*	3.4			
FO *G. miocenica*	3.81			
FO *G. crassaformis viola*	4.0		4.18	
FO *G. exilis*	4.5			
FO *G. crassaformis hessi*	4.5		4.51	
FO *G. crassaformis ronda*	4.5		4.51	
LO *G. nepenthes*	4.88		4.51	
FO *G. puncticulata*	4.88		4.54	
FO *G. crassaformis*	5.06		4.62	
FO *G. margaritae*	6.21		6.13	6.31

Table 2. First (FO) and last occurrence (LO) datums of nannofossils

Datum events	Age (Ma)			
	Site 397	Site 608	Site 609	Site 611
LO *Pseudoemiliania lacunosa*	0.44	0.45	0.45	0.47
LO small *Gephyrocapsa*	0.95			
LO *Helicosphaera sellii*			1.38	1.25
LO *Calcidiscus macintyrei*	1.56		1.55	
FO *Gephyrocapsa oceanica* s.l.	1.87			
LO *Discoaster brouweri*	1.95			
LO *Discoaster pentaradiatus*	2.43		2.43	
LO *Discoaster surculus*	2.57		2.56	
LO *Discoaster tamalis*	2.84	2.76	2.85	
FO *Pseudoemiliania lacunosa*	3.78			
LO *Amaurolithus tricorniculatus*	4.47			
LO *Ceratolithus acutus*	4.70			
LO *Triquetrorhabdulus rugosus*	5.56			
LO *Discoaster quinqueramus*	5.94		5.93	

approximately located at the top of the Mammoth Subchron of the Gauss Chron. The zone is subdivided into the *Globorotalia margaritae margaritae* and *Globorotalia margaritae evoluta* subzones. The base of the *Globorotalia crassaformis* Beds distinguished in the upper part of the latter is defined within the Thvera Subchron of the Gilbert Chron (Table 1).

Within the foraminiferal *Globorotalia margaritae* Zone, the following nannofossil zonation is recognized (Fig. 2).

The uppermost part of the *Discoaster quinqueramus* Zone (CN9) is traceable up to Sample 397-43-2, 51–55 cm, where the index species LO is recorded. This level is 5.94 Ma old (Table 2). The *Triquetrorhabdulus rugosus* Zone (CN10a) corresponding to the lower part of the *Amaurolithus tricorniculatus* Zone NN12 in the standard zonation by Martini spans the interval from the *Discoaster quinqueramus* LO to the index species *Tr. rugosus* LO at 5.56 Ma. The *Ceratolithus acutus* Subzone (CN10b) corresponding to the upper part of the *Amaurolithus tricorniculatus* Zone NN12 is defined as the interval from the *Tr. rugosus* LO to the *C. rugosus* FO. In the hole under discussion, the top of the subzone is placed at the *C. acutus* LO. According to our conclusion, this event took place 4.70 Ma ago. The *Globorotalia crassaformis* FO is recorded within the last subzone at 5.06 Ma. The *Ceratolithus rugosus* Subzone (CN10c), spanning the joint interval of *Ceratolithus rugosus* (NN13) and *Discoaster asymmetricus* (NN14) zones in zonation by Martini, is defined in the section between the *C. acutus* LO and the *Amaurolithus tricorniculatus* LO at 4.47 Ma. The *Reticulofenestra pseudoumbilica* Zone (CN11a, b) corresponds to the Zone NN15 of the Martini's zonation and corresponds to the interval between the *A. tricorniculatus* LO in Sample 397-34-5, 46–50 cm, and the disappearance level of *R. pseudoumbilica* and/or of the genus *Sphenolithus*. Following Bukry, we recognized in Hole 397 the *Sphenolithus neoabies* (CN11a) and *Discoaster asymmetricus* (CN11b) subzones. The boundary between them is marked by the flourishing level of *D. asymmetricus* at the top of Core 30. The *Pseudoemiliania lacunosa* FO in association with small *Gephyrocapsa* forms at 3.78 Ma is the main event within the *R. pseudoumbilica* Zone. The top of the *Discoaster asymmetricus* Subzone at Site 397 is recorded in the middle of the Gauss Chron at 3.20 Ma. Like at Sites 608 and 609, the top of the *Reticulofenestra pseudoumbilica* Zone (CN11) is slightly above the top of the foraminiferal *Globorotalia margaritae* Zone. At Site 397, the latter is 3.23 Ma old and coincides with the top of the Mammoth Subchron of the Gauss Chron.

As noted previously (Bylinskaya *et al.*, 2002), the overlying foraminiferal *Globorotalia miocenica* Zone of subtropical sections can be subdivided into four intervals (faunal beds). In the sediments of Site 397, three lower units among them mainly correspond to the greater part of the *Discoaster tamalis* Subzone (CN12a), which is defined as the interval from the *R.*

pseudoumbilica LO to the *D. tamalis* LO at 2.84 Ma. The upper *Globorotalia exilis* Beds correspond to three nannofossil subzones distinguishable in the lower part of the *Discoaster surculus* Zone NN16 of the Martini scale. The lower *Discoaster surculus* Subzone (CN12b) is established between the *D. tamalis* and *D. surculus* extinction events, the latter recorded in the lowermost Matuyama Chron at 2.57 Ma. The next *Discoaster pentaradiatus* Subzone (CN12c) with the top at 2.43 Ma corresponds to the Zone NN17 by Martini, and the third unit spans (in this hole) the lower part of the *Discoaster brouweri* Subzone (Fig. 2). As is evident, the established datums are very close to that cited in the publications (Kučera, 1998; Kameo and Takayama, 1999, Spencer-Cervato, 1999). Thus, different groups of calcareous plankton from Site 397 enable recognition of 5 to 6 biostratigraphic units in the time span from 3.3 to 2.2 Ma.

It should be noted that certain biostratigraphic boundaries recognized at Site 397 differ in age from their analogues established in other holes (see below) and known from the literature. As reported previously (Bylinskaya, 1999), the southern part of the subtropical belt is characterized by the earlier appearance and later extinction of some calcareous plankton species. For instance, the published (Berggren *et al.*, 1995) and our own data on other holes (see below), the first occurrence levels of *Globorotalia crassaformis, G. puncticulata, G. miocenica*, and *Gephyrocapsa oceanica* s.l. are lower at Site 397, whereas the last occurrences of *Globorotalia margaritae* and, likely, of *Sphenolithus* forms are recorded here considerably higher (Tables 1 and 2).

The foraminiferal *Globorotalia tosaensis* Zone of the late Pliocene is defined within the interval from the *G. exilis* LO (*G. miocenica* disappeared slightly earlier at the site) to the *G. truncatulinoides* FO at 1.95 Ma. The zone corresponds to the last *Discoaster brouweri* (=*Calcidiscus macintyrei*) Zone CN12d of Pliocene nannofossils, the analogue of the Zone NN18 by Martini, and its top corresponds to the last occurrence level of index species at the base of the Olduvai Subchron. At Site 397, this event coincides with the appearance of *Globorotalia truncatulinoides*. Therefore the Pliocene–Quaternary boundary in this hole is identical in zonations of both the planktonic foraminifers and nannofossils.

In the Quaternary interval, the foraminiferal *Globorotalia truncatulinoides* Zone and five subzones after Bolli and Saunders (1985) are distinguished. Subdividing sediments based on nannoplankton, we used the Gartner's zonation and recognized in addition the zones after Okada and Bukry. Impediments for a detailed subdivision were therewith a turbidite layer present in Core 12 and an evident hiatus in the *Helicosphaera sellii*–small *Gephyrocapsa* interval. The foraminiferal *Globorotalia crassaformis hessi* Subzone shown in Fig. 2 is strongly reduced because of the hiatus.

In the lower Quaternary, two nannofossil zones and two subzones of planktonic foraminifers are distinguished at the site. The latter ones are the *Globorotalia crassaformis viola* and *G. crassaformis hessi* subzones, the last one strongly reduced in this hole. The *Calcidiscus macintyrei* Zone by Gartner is defined between the *D. brouweri* and *C. macintyrei* LO levels, and its top 1.56 Ma old is recorded in Sample 397-12-2, 62–66 cm. The *G. crassaformis viola* LO cannot be determined exactly because of the hiatus in the section. When correlating previously the calcareous planktonic zonations in the tropical Atlantic (Bylinskaya and Golovina, 1990), we also placed the top of the *C. macintyrei* Zone within the *G. crassaformis viola* Subzone.

The most important event in the interval under consideration is the first occurrence of typical large *Gephyrocapsa oceanica* s.l. in Sample 397-14-4, 60-64. Their first occurrence is dated back to 1.87 Ma, and this level is the oldest one among dates reported previously. The interval between the *D. brouweri* LO and *G. oceanica* s.l. FO corresponds to the Zone CN13 in the low-latitude zonation elaborated by Okada and Bukry.

Upward the section, the undivided *Helicosphaera sellii*–small *Gephyrocapsa* interval corresponds to the foraminiferal *Globorotalia crassaformis hessi* Subzone coupled with conjoined parts of the over- and underlying subzones. In studied section, the top of the *Gephyrocapsa* Zone is placed by convention above the upper boundary of the *Globorotalia crassaformis hessi* Subzone.

In the upper Quaternary interval of the section, we distinguished three foraminiferal and three nannofossil units. The *Pseudoemiliania lacunosa* Zone is correlated with a largest part of the foraminiferal *Globigerina calida calida* Subzone; its top at the level of 0.44 Ma is recorded within the *Globigerina calida calida* Subzone like in the tropical Atlantic. The upper boundary of the *Gephyrocapsa oceanica* Zone is defined by the first appearance of *Emiliania huxleyi* in Sample 397-2-5, 40–41 cm. As in the tropical Atlantic, the zone is within the *Globigerina calida calida* Subzone. At Site 397, the base of the overlying *Emiliania huxleyi* Zone is placed slightly below the base of the foraminiferal *Globigerina bermudezi* Subzone. The *Emiliania huxleyi* Acme zone was not distinguished in the studied cores.

Site 608

This DSDP site is located considerably northward of Site 397. However, zonations of planktonic foraminifers and nannofossils at both sites are well correlative, as we shall see below (Fig. 3).

We studied the calcareous plankton from the Pliocene–Quaternary interval starting with the middle part of Core 17. The Lower Pliocene *Globorotalia mar-*

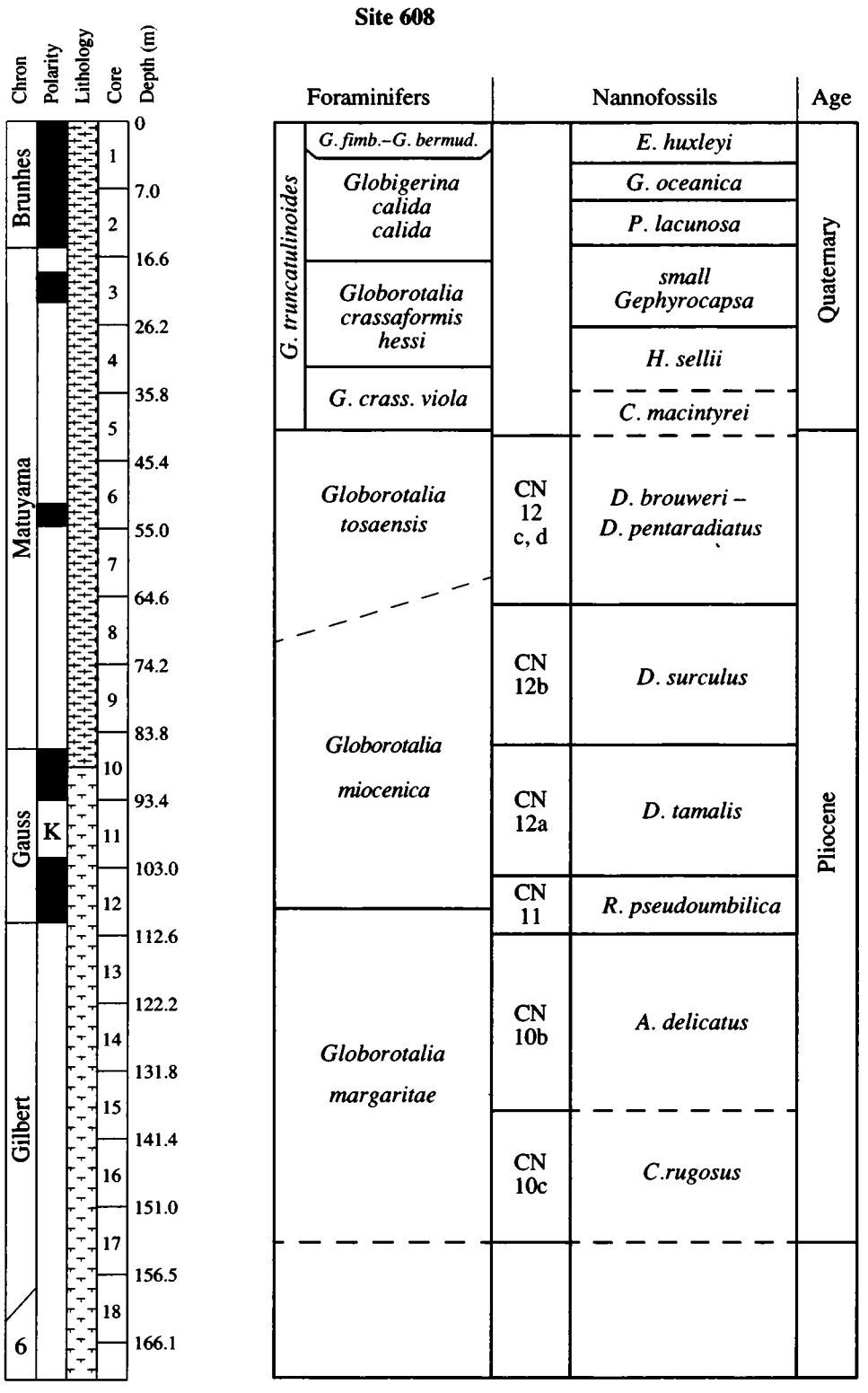

Fig. 3. Pliocene–Quaternary zonal units of calcareous plankton from DSDP Site 608 (symbols as in Fig. 2).

garitae Zone corresponds at the site to successive nannofossil zonal units. The lower *Ceratolithus rugosus* Subzone CN10c extends upward to the basal interval of Core 15, Sample 608-15-5, 95–97 cm, inclusive. The first occurrence of planktonic foraminifer *Globorotalia crassaformis* is recorded within the subzone, and its top coincides with the LO levels of *Globorotalia conomiozea* and *G. conoidea*. Higher in the section up to Sam-

ple 608-13-1, 99–101 cm, we recognized the *Amaurolithus delicatus* Subzone. Within this subzone, there is recorded the first occurrence of planktonic foraminifers *Globorotalia crassaformis viola*, and characteristic of its upper part are the last occurrence of *Globigerina nepenthes* and the appearance of *Globorotalia crassaformis hessi*. However, we failed to determine age of these events, because the magnetochronological records in this interval are unreliable (Weaver and Clement, 1986).

Higher in the section up to Sample 608-12-2, 100–102 cm, the undivided *Reticulofenestra pseudoumbilica* Zone (CN11) is recognized. Its range is evidently reduced. The top of the zone is placed, as at Site 397, in the lowermost foraminiferal *Globorotalia miocenica* Zone.

The undivided interval of *Globorotalia miocenica* and *Globorotalia tosaensis* zones includes here, like at Site 397, the succession of *Discoaster* subzones, of which the lowermost *Discoaster tamalis* Subzone (CN12a) with top dated at 2.76 Ma ranges upward to Sample 608-10-1, 96–98 cm. Planktonic foraminifers *Sphaeroidinellopsis seminulina, Globoquadrina altispira*, and *Globorotalia pertenuis* successively disappear within this subzone as at Site 397 (Table 1). The subzone lower boundary is placed, with certain reservations, at the last occurrence level of *Reticulofenestra pseudoumbilica.*

The overlying interval up to Sample 608-8-2, 95–97 cm, is assigned to the *Discoaster surculus* Subzone (CN12b). The subzone yields comparatively abundant and diverse nannofossils whose assemblage is of a relatively cold-water type, as indicated by abundance of *Coccolithus pelagicus* and by rarity of *Discoaster* forms. The base of the subzone is defined by the last occurrence of index species of the underlying unit. The foraminifer assemblage that includes the first appeared *Globorotalia tosaensis, G. inflata*, and *G. triangula* is also of a rather cryophilic character.

Upward the section the interval of the *Discoaster pentaradiatus* (CN12c) and *Discoaster brouweri* (=*Calcidiscus macintyrei* CN12d) subzones is recognized. It is difficult to separate undoubtedly these two units at Site 608. The top of the latter corresponds to the last occurrence of the index species in Sample 608-5-5, 68–70 cm.

It should be particularly emphasized that Sites 608 and 609 are located in the narrow geographic belt between 42° and 49° N, where the Pliocene–Quaternary boundary can be concurrently defined based on three events: on simultaneous extinction of *Discoaster brouweri* and *Neogloboquadrina atlantica* sin. and on the first appearance of *Globorotalia truncatulinoides.* Northward of the belt, the latter species appears for the first time later, whereas to the south species *Neogloboquadrina atlantica* are missing.

The Quaternary lower boundary that coincides with the base of the foraminiferal *Globorotalia truncatulinoides* Zone, is placed at Site 608 slightly above the *Discoaster brouweri* Subzone, the top of which is defined in Sample 608-5-2, 96–98 cm. The top of the *Calcidiscus macintyrei* Zone is within the foraminiferal *Globorotalia crassaformis viola* Subzone, as in the other studied sections, though at the site in question this level is established conventionally. Upward the section the *Helicosphaera sellii* Zone is distinguished up to the extinction of its index species, and the zone top is placed in Sample 608-4-1, 97–99 cm, within the foraminiferal *Globorotalia crassaformis hessi* Subzone. In the upper part of the zone (Sample 608-4-2, 94–96 cm), the first occurrence of large *Gephyrocapsa oceanica* is recorded.

Zone of small *Gephyrocapsa* forms occupies the uppermost part of the *Globorotalia crassaformis hessi* Subzone and partly overlaps the lowermost foraminiferal *Globigerina calida calida* Subzone. Sample 608-2-6, 95–97 cm, is typical of the zone and bears the nannofossil assemblage completely consisting of the index species. The Jaramillo Subchron is recorded in the middle part of the zone, as at Sites 397 and 609, whereas in the piston cores from tropical Atlantic the zone top corresponds to the subchron lower boundary (Bylinskaya and Golovina, 1990).

The *Globigerina calida calida* Subzone also includes ranges of the *Pseudoemiliania lacunosa* Zone (up to Sample 608-2-2, 98–100 cm) whose upper boundary is at 0.45 Ma and of the *Gephyrocapsa oceanica* Zone (up to Sample 608-1-5, 68–70 cm). The foraminiferal *Globigerina bermudezi* Subzone corresponds to the *Emiliania huxleyi* and *Emiliania huxleyi* acme zones.

Site 609

This site is located in the temperate belt, where the assemblages of foraminifers and nannofossils are substantially impoverished (Bylinskaya *et al.*, 2002). However, almost complete succession of zonal units of both groups of calcareous plankton can be recognized here (Fig. 4), and the units are close to those established at Site 608 (Fig. 3).

The *Globorotalia margaritae* Zone is distinguishable in the section beginning from the uppermost part of Core 37. Its base almost coincides in age with the same datum at Site 397 (Table 1). The range of the zone includes the following nannofossil units.

The upper part of the *Discoaster quinqueramus* Zone (CN9) is recovered in the lowermost Pliocene interval up to Sample 609-36-2, 98–100 cm. The last occurrence level of index species at 5.93 Ma is close to the same level at Site 397. A series of thermophilic species present in the assemblage permits the recognition of the interval as the *Amaurolithus primus* Subzone CN9b. The *Amaurolithus tricorniculatus* Zone CN10 undivided into subzones is distinguishable up to Sample 609-28-2, 99–101 cm. The first occurrence of

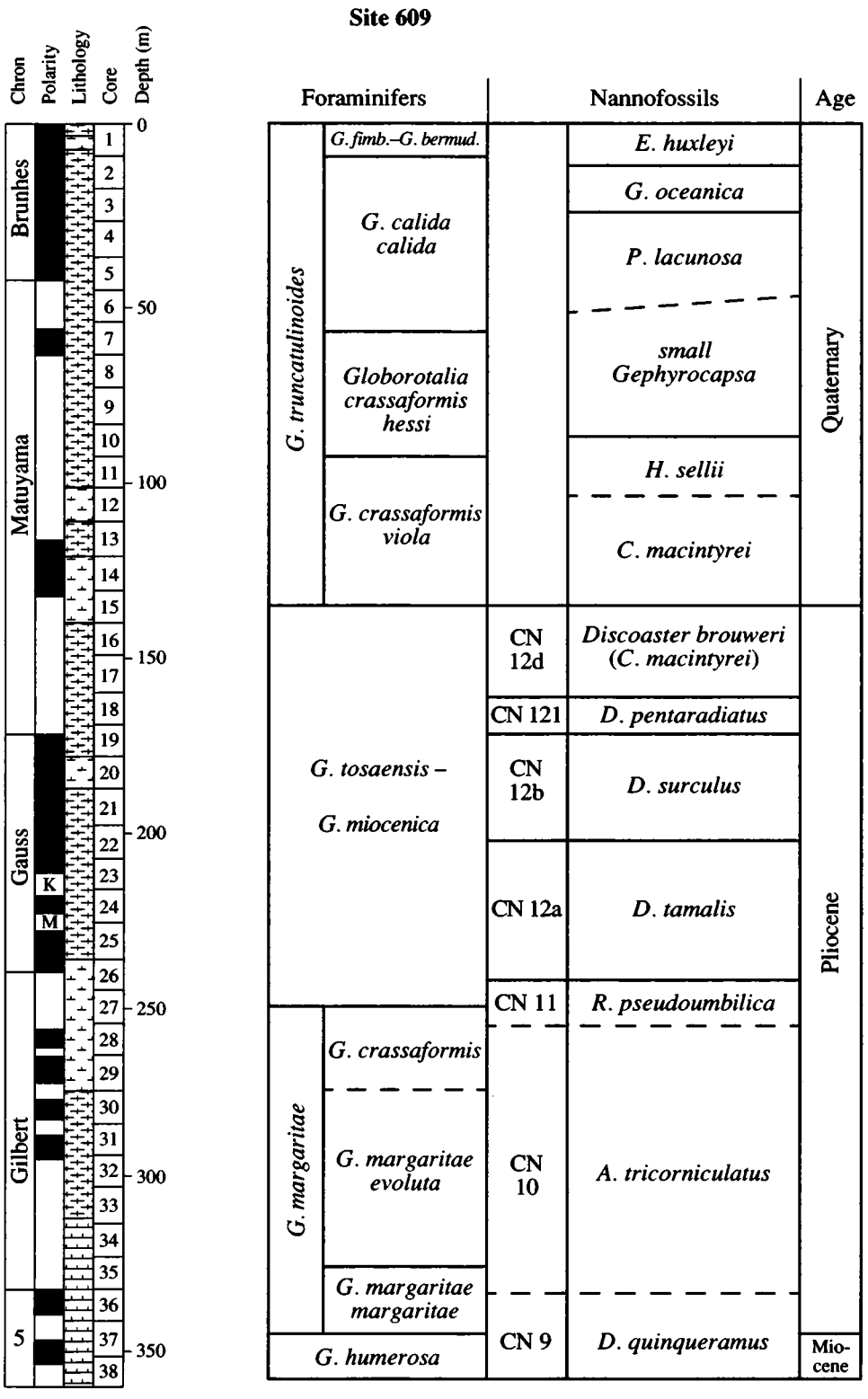

Fig. 4. Pliocene–Quaternary zonal units of calcareous plankton from DSDP Site 609 (symbols as in Fig. 2).

planktonic foraminifers *Globorotalia crassaformis* and *G. puncticulata* is recorded within the zone. The *Reticulofenestra pseudoumbilica* Zone (CN11) is likely reduced at the site (Fig. 4). Like at Site 608, its top is defined in Sample 609-26-6, 98–100 cm, from the lowermost foraminiferal *Globorotalia miocenica* Zone.

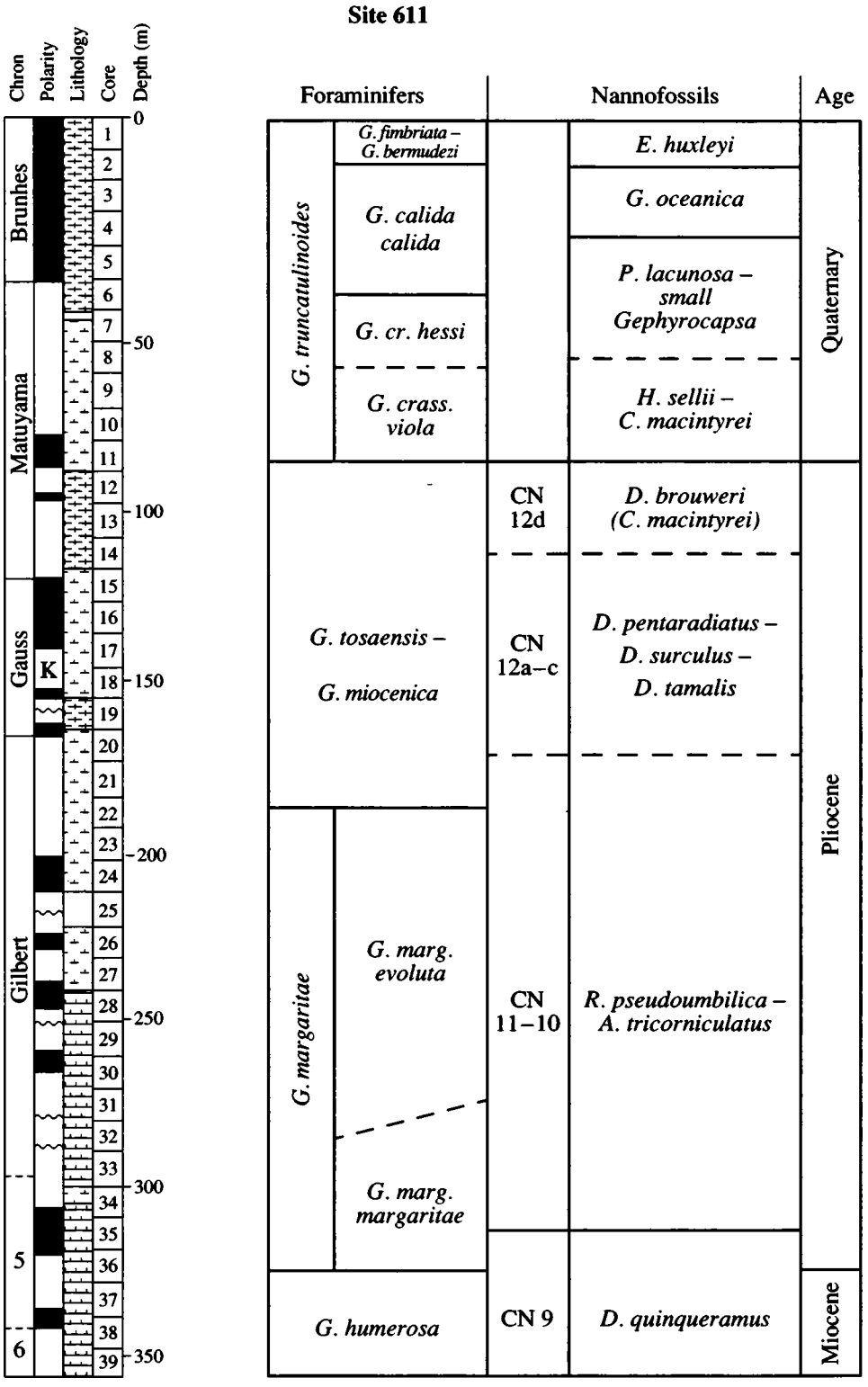

Fig. 5. Pliocene–Quaternary zonal units of calcareous plankton from the DSDP Site 611 (symbols as in Fig. 2).

The last occurrence of *Globorotalia margaritae* recorded within the zone took place at this northern site considerably earlier than in the low-latitude regions.

As in the other studied sections, the undivided interval of the *Globorotalia miocenica* and *Globorotalia tosaensis* foraminiferal zones corresponds to several

nannofossil units. The *Discoaster tamalis* Subzone ranges up to Sample 609-22-2, 57–59 cm, and its top is dated at 2.85 Ma. The *Discoaster surculus* Subzone extends up to Sample 609-19-1, 98–100 cm (age level of 2.56 Ma) and the *Discoaster pentaradiatus* Subzone up to Sample 609-18-1, 98–100 cm (age level of 2.43 Ma). The *Discoaster brouweri* (=*Calcidiscus macintyrei*) Subzone CN12d terminates the Pliocene section at Site 609. Its upper boundary in Sample 609-15-4, 103–105 cm, is very close to the extinction level of *Neogloboquadrina atlantica* and to the *Globorotalia truncatulinoides* FO.

As at Site 608, the *Calcidiscus macintyrei* Zone of the lower Quaternary corresponds to a greater part of the *Globorotalia crassaformis viola* Subzone. Its top in Sample 609-12-1, 102–104 cm, is dated back to 1.55 Ma. The *Helicosphaera sellii* Zone is overlapping with the lower half of the *Globorotalia crassaformis hessi* Subzone. The top of the former defined in Sample 609-10-3, 102–104 cm, is 1.38 Ma old. Despite the conventional upper boundary of the *Gephyrocapsa* Zone at the site, it seems to be close to the top of the *Globorotalia crassaformis hessi* Subzone. The *Globigerina calida calida* Subzone includes the *Pseudoemiliania lacunosa* Zone ranging up to Sample 609-3-6, 96–98 cm (age level of 0.45 Ma), and the *Gephyrocapsa oceanica* Zone traceable up to Sample 609-2-3, 103–105 cm. Finally, the undivided interval of the *Globigerina bermudezi* and *Globigerina fimbriata* subzones corresponds to the *Emiliania huxleyi* Zone.

Site 611

DSDP Site 611 located in the Boreal Atlantic is the northernmost one among the others. In distinction from other cases, when we were able to correlate in detail the zonal units established for different groups of calcareous plankton, the microplankton assemblages from Site 611 are so impoverished that it was impossible to divide the recognizable zones and intervals into smaller units (Fig. 5).

The *Globorotalia margaritae* Zone is correlated at the site with the upper part of the nannofossil *Discoaster quinqueramus* Zone CN9 whose top is in the uppermost Chron 5, like at Sites 609 and 397, and with the greater part of zones CN10 and CN11. The upper boundary of the undivided *Amaurolithus tricorniculatus* (CN10)–*Reticulofenestra pseudoumbilica* (CN11) interval is established in the upper part of the Gilbert Chron within the interval of the foraminiferal *Globorotalia miocenica–Globorotalia tosaensis* Zone. This level analogous to its counterpart at Site 609 is dated back to 3.63 Ma based on the last occurrence of *Sphenolithus* forms. It is remarkable that the estimated disappearance time of *Sphenolithus* forms at this northern site coincides with dates established in other climatic belts (Berggren *et al.*, 1995). The rest of the *Globorotalia miocenica–Globorotalia tosaensis* interval includes the *Discoaster* subzones, first the undivided interval of the *Discoaster tamalis* (CN12a), *Discoaster surculus* (CN12b), and *Discoaster pentaradiatus* (CN12c) subzones ranging up to Sample 611-14-3, 101–103 cm, and then the *Discoaster brouweri* (=*Calcidiscus macintyrei*) Subzone (CN12d) ranging up to Sample 611-11-5, 99–101 cm. The top of the latter subzone coincides with the Pliocene–Quaternary boundary that was defined in the studied section at the last occurrence level of *Neogloboquadrina atlantica* and corresponds to the Olduvai Subchron.

The units recognized in the Quaternary part of the section are as follows. The *Globorotalia crassaformis viola* Subzone approximately corresponds to the greater part of the *Calcidiscus macintyrei–Helicosphaera sellii* undivided interval, the top of which conventionally defined in Sample 611-8-3, 103–105 cm, is dated back to 1.25 Ma. The *Gephyrocapsa–Pseudoemiliania lacunosa* undivided interval ranging up to Sample 611-4-5, 98–100 cm, includes the *Globorotalia crassaformis hessi* and, partly, *Globigerina calida calida* subzones. Crowning the section are the *Gephyrocapsa oceanica* and *Emiliania huxleyi* nannofossil zones.

CONCLUSIONS

The analyzed correlation between nannofossil and foraminiferal zones in the Pliocene–Quaternary deposits of subtropical and boreal region of the Atlantic Ocean leads to the following conclusions.

(1) The upper part of the *Discoaster quinqueramus* Zone (CN9) in all studied sections is overlapping with the Lower Pliocene foraminiferal *Globorotalia margaritae* Zone, and its top is close to the base of the Gilbert Chron. The base of the *Globorotalia margaritae* Zone is 6.1 to 6.3 Ma old in the sections, and the top of the *Discoaster quinqueramus* Zone is 5.94 Ma old.

(2) The foraminiferal *Globorotalia margaritae* Zone (N18–N19) is comparable in range with nannofossil zones CN10–CN11 and includes the *Triquetrorhabdulus rugosus* (CN10a), *Ceratolithus acutus* (CN10b), *Ceratolithus rugosus* (CN10c), *Amaurolithus delicatus* (CN10d), *Sphenolithus neoabies* (CN11a), and *Discoaster asymmetricus* (CN11b) subzones after Okada and Bukry (1980). In all studied sections, the top of the *Reticulofenestra pseudoumbilica* Zone (CN11) is established in the lower part of the foraminiferal *Globorotalia miocenica* Zone close to the boundary between the Gilbert and Gauss chrons.

(3) The greater part of the foraminiferal *Globorotalia miocenica* and *Globorotalia tosaensis* (N20–N21) zones corresponds to a succession of *Discoaster tamalis, Discoaster surculus, Discoaster pentaradiatus* and *Discoaster brouweri* (=*Calcidiscus macintyrei*) subzones CN12a–CN12d.

(4) The lower boundary of the foraminiferal *Globorotalia truncatulinoides* Zone corresponds to the base of the Olduvai Subchron and almost coincides

with the top of the *Discoaster brouweri* Subzone at the level of 1.95–2.0 Ma according to our assessment. The Pliocene–Quaternary boundary is thus substantiated based on distribution of nannofossils and planktonic foraminifers. In a narrow belt of the Atlantic Ocean between 42° and 49° N, the Pliocene–Quaternary boundary marks three concurrent events: the extinction of *Discoaster brouweri* and *Neogloboquadrina atlantica* sin and the first appearance of *Globorotalia truncatulinoides*. To the north of the belt, the first occurrence level of latter species is recorded higher, whereas *Neogloboquadrina atlantica* is missing to the south of the belt.

(5) In the Quaternary sequences of all studied sites, similar correlative zonations of two groups of calcareous plankton are observable. The *Calcidiscus macintyrei* Zone (Gartner, 1977) is correlative in this case with the lower (greater) part of the *Globorotalia crassaformis viola* Subzone. Zones of *Helicosphaera sellii* and small *Gephyrocapsa* forms approximately correspond to the *Globorotalia crassaformis hessi* Subzone, and the Jaramillo Subchron is therewith in the second nannofossil zone. The *Pseudoemiliania lacunosa* Zone is included in or overlaps with the *Globorotalia calida calida* Subzone. The *Gephyrocapsa oceanica* and *Emiliania huxleyi* zones correspond to the upper part of the *Globorotalia calida calida* and two overlying subzones.

ACKNOWLEDGMENTS

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Viktor Efimovich Khain (on 90th Birthday)

Academician of the Russian Academy of Sciences Dr. Viktor Efimovich Khain, a member of our editorial board, senior scientists at the Institute of Lithosphere of Marginal Seas RAS, honored professor of the Moscow State University and worldwide known scientist, the acknowledged authority in tectonics, met his 90th birthday.

Viktor Efimovich was born February 26, 1914, in Baku. In 1935, he graduated from the Industrial Institute of Azerbaijan and started his scientific career as a petroleum geologist.

The results of long-lasted scientific studies by V.E. Khain are published in more than 1200 works, 800 of which, 43 monographs included, are fundamental, highly valued by geologists. Over 70 of his works are published abroad in English, French, German, Spanish, Chinese, Polish and Romanian languages. His last monographs issued at the Institute of Lithosphere of Marginal Seas RAS and entitled "The tectonics of Continents and Oceans" (2001) and "Basic Problems of Contemporaneous Geology" (2003) are of extremely high theoretical value. V.E. Khain always combined his scientific activity with the educational work and still lectures at the Moscow State University, where he was acknowledged as an outstanding professor in 2003.

Academician V.E. Khain has many honorable titles. He is a fellow of many national and international scientific societies: the Russian Academy of Natural Sciences, European Academy, Georgian Academy of Science, Royal Society of London, American, French and Bulgarian geological societies, Moscow Naturalist Society, and Russian Association of Petroleum geologists. He is also an Honorable doctor of Pierre and Marie Curie University of Paris and an Honorable professor of the Xiling University (China).

Among the valuable scientific rewards of V.E. Khain, there are the State prizes of the USSR and the Russian Federation, the Lomonosov premium of the Moscow State University, three prizes of the Moscow Naturalist Society, and the Arkhangelsky prize and Karpinsky gold medal of the Russian Academy of Sciences. He is recognized with the award of gold medals by the P. Fourmarier Royal Academy of Belgium and by the National Economic Expo of Russia. He is also the Prestvitch prizewinner of the French Geological Society and the Steinmann prizeman of the German Geological Society.

V.E. Khain is a national treasure of Russia and one of the most famous geologists of the world. Being a person of encyclopedic knowledge, he bears in harmony the talents of geologist, orator and organizer. His bright aura is beacon of the wise for young geologist, his followers, 200 of which defended candidate's and doctorial dissertations under his supervision.

The authority of V.E. Khain, his fidelity to science and ability to discuss attentively the controversial opinions and fact, to help with advice and knowledge, create an extraordinary warm atmosphere around him that attracted geologists of many generations. We wish to Victor Efimovich the all joys of life and many years of fruitful work.

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